Evolution of a late triassic fluvial system, NW Libya

Mayouf, Giuma M.

DURHAM UNIVERSITY

EVOLUTION OF A LATE TRIASSIC FLUVIAL SYSTEM, NW LIBYA

Giuma M. Mayouf

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Supervisor: Dr. Stuart J. Jones

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The Department of Earth Sciences
Durham University

Declaration

I declare that the work contained in this thesis was the result of my independent research except where otherwise stated.

Signed ______________________

Giuma Musa Mayouf

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ABSTRACT

Fluvial systems, channel sandbodies and valley fills have been widely studied and documented in the literature, with particular attention paid to their internal structures. Although fluvial architecture and channel body geometry have been widely linked to baselevel change, tectonics and accommodation space, few studies have taken a holistic approach of additionally incorporating climatic signatures, diagenesis and width/thickness (W/T) of low net to gross fluvial systems. To remedy this situation, one of the first studies of a series of excellent fluvial exposures of the Late Triassic Abu Shaybah Formation (216.5 -199.6 Ma; c.15.9 Myrs), located along the Jabal Nafusah of northern Libya have been examined.

The fluvial deposits of the Abu Shaybah Formation have been recognized to be part of an incised valley fill situated along a major strike slip fault zone, now demarked by the Jabal Nafusah escarpment. The associated architecture of the fluvial systems are identified by channel sand bodies with low W/T <10 for the valley fills, due to incision along tectonic linements and stacking along faults. In comparison passing up section away from the lateral confinement the sand bodies develop into mobile channel belts of braided and meandering stream systems with W/T >10-15. The architecture of the Abu Shaybah Formation changes up section as it becomes less influenced by tectonic activity, while climatic signatures become more marked by variables in stream discharges and sediment flux, recognized by channel aggradation and degradation of the fluvial systems. Such an occurrence has been used to help correlate the fluvial succession laterally and recognise how tectonism provides the space for and distribution of rivers while climate controls the sediment supply and facies evolution.

Detailed petrographic analysis of the Abu Shaybah Formation has revealed the importance of new clay mineral growths for the identification of sequence boundaries.

This research demonstrates the importance of taking a holistic approach when studying fluvial successions and has important implications for low net to gross fluvial reservoirs as a potential exploration target in the Al Kufrah and Sirte Basins.

Contents

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Chapter 1
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1.1. Introduction

The collision between Gondwana and Laurasia resulted in the formation of a supercontinent which included virtually all of the world's continental plates. This supercontinent, which was named Pangaea by Wegener in 1966, and persisted until the Jurassic. The surrounding World ocean was called Panthalassa. The collision involved a significant strike-slip component, and a major dextral shear-zone developed along the line of contact. The configuration of several of the Triassic basins along the North African margin is controlled by the pull-apart geometry propagated by the shear-zone. Break-up of Pangaea began in the Triassic. In Appalachia rift grabens are filled with late Triassic clastics, and in North Africa there is evidence of Triassic rifting in Tunisia and Libya. The North Africa continental margin at this time was characterised by extension and crustal thinning. In Algeria a blanket of Triassic fluvial sands and shales was deposited on the eroded Palaeozoic surface and these sands are overlain by extensive evaporites, which in turn are overlain by shelf carbonates. This sequence illustrates a progressive deepening of the sea due to post-rift thermal subsidence. Break-apart also began in the late Triassic between the former components of East and West Gondwana, with rifting along the southern margin of the Arabian plate, and along the eastern margin of Africa.

The break-up of Pangaea involved the establishment of a spreading axis throughout the Mediterranean region. Associated igneous activity in Libya has been reported from the Waddan Platform (granodiorite 230 Ma) and from the Amal oilfield (granodiorite 207 Ma). These events correspond to the initial rifting phase in the Mediterranean region. Tensional faults of Triassic age are present both offshore and onshore in Libya, and several unconformities are present within the Triassic succession. Evidence from eastern Libya shows the presence of Triassic sediments which may represent the earliest deposits in incipient syn-rift grabens. Triassic rocks are present in offshore Libyan basins and thin onto the Jabal Nafusah Uplift which again marks the effective Tethyan shoreline. Over the rest of Libya Triassic rocks are almost entirely continental in character.

During the Triassic, the UK lay beyond the Western termination of the Tethys Ocean, which was divided into a northern part, the Palaeotethys, and a southern part, the Neotethys (Fig. 1.1) Between these oceans occurred the Cimmerian terrains; several now widely separated continental fragments which had rifted from the northern fringe of Gondwana in the Permian (Stampfli & Borel 2002). The Triassic witnessed the northward drift of these Cimmertian terrains, and the northward subduction of the Palaeotethys, which was mostly completed by the Late Triassic. However, in the western Tethys, some Triassic ocean basin fragments (e.g. Pindos, Meliata) persisted into the Jurassic (Fig.1.1Stampfli & Borel 2002). The southern margin of the Neotethys Ocean stretched southeast wards through North Africa, Arabia, northern India and NW Australia and shows clastic and carbonate Triassic successions typical of a passive margin contrasting strongly with the destructive northern margin of Palaeotethys (Weidlich & Bernecker 2003).

Extensive rift basins formed in Europe during the Triassic, stretching from the Boreal Ocean in the north, through the Greenland-Norwegian gateway into many of the bordering regions around Britain and Ireland, and extending south and SE into France, Iberia and central Europe (Fig. 1.1). This rifting extension was a continuation of Permian rifting and segmented both Caledonide and Variscan basement.

1.2 Aims

The thesis investigates a series of Late Triassic outcrops along the Jabal Nafusah, NW Libya. This area not only affords excellent exposures of Late Triassic fluvial red beds but allows for detailed correlation and recognition of climatic and tectonic signatures in the sedimentary succession. The research is primarily focused on four main aims:

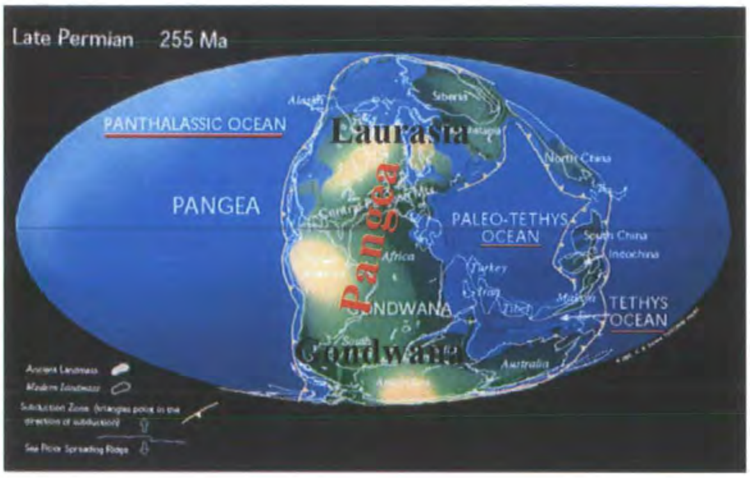
- Investigation of fluvial architecture in a incised valley fill and the application of sequence stratigraphy to non-marine strata;
- Identifying tectonic and climatic signatures in the Late Triassic Abu Shaybah Formation and how these can be used for correlation purposes;

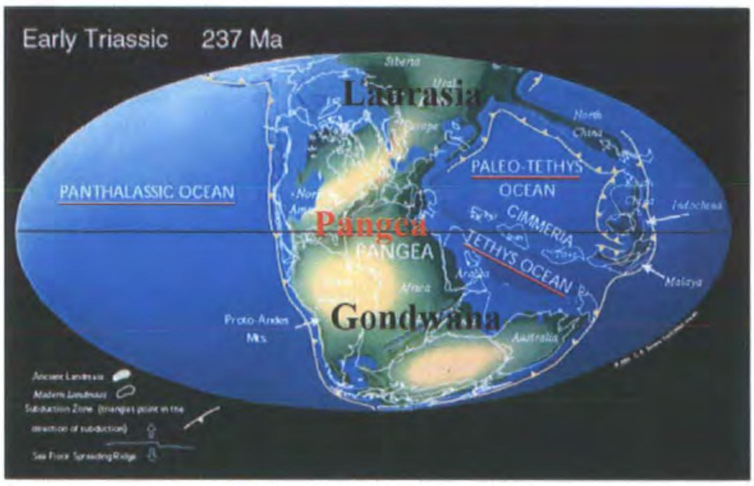
- Investigation of the petrography and diagenesis of the Abu Shaybah Formation for improved reservoir understanding of fluvial successions and the application of sequence stratigraphy through the use of diagenetic studies and;
- Improved understanding of low-net to gross reservoirs and comparison to other important Triassic petroleum plays.

1.3. Palaeogeography and signatures in Europe and North Africa

The Permian witnessed the fragmentation of the core of the Variscan Mountains into a number of separate areas, such as the Armorican, Bohemian and Iberian massifs (Fig. 1.1). These supplied sediment northwards into adjacent basins during much of Early and Mid Triassic time. It was only marginal to the Fennoscandia uplands, and during parts of the Carnian (Late Triassic) in Central Europe, that sediment sources switched to the north or northeast (Ziegler 1990).

The south polar glaciations, which influenced the Permian, were not a factor in the Triassic, with no evidence of polar ice caps. The Pangaea supercontinent during the Triassic severely perturbed the latitudinal changes in palaeoclimate (Parrish 1993). The interior northern regions of the supercontinent would have experienced low-pressure conditions in the summer, drawing humid air from the Tethys Ocean. Much of this moisture may have precipitated on the remnant Variscan mountains in southern-central Europe and Iberia, leading to northwards seasonal transport of sediment by major river systems. In the winter, the northern interiors would have been sources of dry, externally directed winds. This strongly seasonal monsoonal climate was its peak during the Triassic and was enhanced by the equatorial position of Tethys, flanked to the north and south by large continental masses (Parrish 1993). Cross – stratification measurements on dune systems in the upper parts of the Sherwood Sandstone Group (Middle Triassic) mostly demonstrate that they were deposited by winds blowing from the Triassic ENE to NNE (Thompson 1970b; Meadows & Beach 1993; Jones & Ambrose 1994; Mountney & Thompson 2002). These may have been the monsoonal dry winter winds emanating from the interior of northern Pangaea. This monsoonal climate would give a strong seasonality,





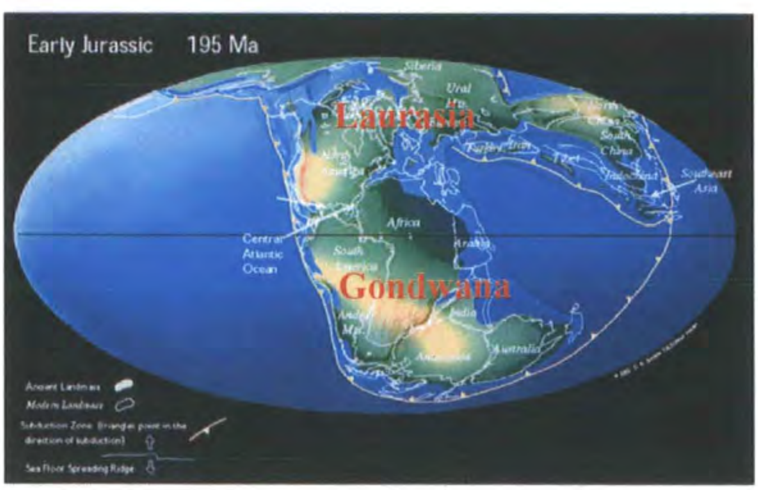


Fig. 1.1 Paleogeographic of the world in the Early Jurassic, Early Triassic and Late Permian, by C.R. Scotese, PALEOMAP Project (www.scotese.com).

possibly even at the equator (Parrish 1993). Consequently, even though England and Wales drifted northwards during the Triassic from a palaeolatitude of about 16°N – 34° N, this drift may not have radically altered the climate, which was dominated by strong monsoonal circulation. Interior parts of Pangaea may have experienced arid and semi-arid conditions, stretching from near the equator to around 50° N, and perhaps extended over 80% of Pangaea in the Early Triassic (Chumakov & Zharkov 2004).

During the Triassic much of southern Europe (south of the Alps and Caucasus Mountains) comprised a series of fault-bounded blocks at the SW termination of the Tethys Ocean (Fig.1.1). The Mid and Late Triassic saw these blocks fragment, and subside, producing fringing and ramp carbonate reefs with intervening deep marine basins.

Within a zone across central Europe (Poland-Germany-Denmark-France-England-Wales- Spain), Triassic basinal successions can be broadly divided into a Lower Triassic fluvial sandstone, red-bed dominated succession, overlain by Middle and Upper Triassic succession that is dominated by mudstones, carbonates and evaporites. The extent to which this reflects palaeoclimatic or eustatic sea-level changes, end-Permian biotic recovery or regional changes in the tectonic stress regime is debatable (Ziegler 1990; Jacouin & de Graciansky 1998; Benton & Twitchett 2003). The timing of this switch, at the transition from the Lower to Middle Triassic in the red bed succession of central Europe, also coincides with changes in the depositional style in many Tethyan successions on the southern European margin (Ruffer & Zuhike 1995; Gianolla & Jcquin 1998). Central European areas that were adjacent to local clastic sources, such as the fringes of Fennoscandia, or the dissected fragments of the Variscan Mountains, may not show the same pattern and timing of changes.

The red-bed Triassic successions in Europe were characteristically deposited in graben and extensional basins, with a variety of orientations, which may reflect both underlying basement structures and regional stress field (Peacock 2004). The relative timing of basin formation and filling was quite variable. Locally an Early Triassic formation and filling phase occurred in areas such as the Irish and Scottish Atlantic margins and northern North Sea. Major extension does not appear to have taken place in many European areas until Mid or Late Triassic times (Zeigler 1990), as in many graben systems in North Africa and the eastern seaboard of North America.

The late stage of the Triassic (the Rhaetian) witnessed marine flooding of many of the continental basins of Europe, with the exception of some in Europe and those fringing Fennoscandia. This pre-empted the marine connection of the Tethys and Boreal Oceans in the Jurassic through the Greenland-Norway rift system.

1.4. Stratigraphic subdivision

1.4.1. Triassic in Libya – European equivalents (Fig. 1.2)

The Triassic, as originally proposed by Friedrich August von Alberti in 1834, comprised three lithostratigraphic subdivisions based on the German succession, the Bunter, Muschelkalk and Keuper. The Triassic Period is now subdivided into seven stages (Induan, Olenekian, Anisian, Ladinian, Carnian, Norian and Rhaetian).

The Triassic section in Jabal Nafusah (NW Libya) is similar, even though it may not be identical to the European (Germanic Facies) succession (Fig. 2.1) which consists mainly of i) basal Early Triassic continental dominated siliciclastic red beds related to deposition in rivers and lakes; ii) Middle Triassic carbonate section called Muschelkalk that consists of brackish lagoonal deposits, which at the base changing into limestone and marls followed by coal bearing sandstones and marls at the later stage ; and iii) a Late Triassic red beds associated with lagoonal marls, evaporites and even some coal. This succession is broadly equivalent to Kurrush (Ras Hamia) Formation clastics, Al Aziziyah carbonates and Abu Shaybah clastics respectively. It is reasonable to assume that the middle limestone unit (Muschelkalk - Al Aziziyah) represents a major mid-Triassic eustatic sea level rise in the Pangea. In the UK, the succession consists of only two units. A lower continental red bed called the Sherwood sandstone group followed by a with extensive deposits of salt and mudstones (Fig. 1.2).

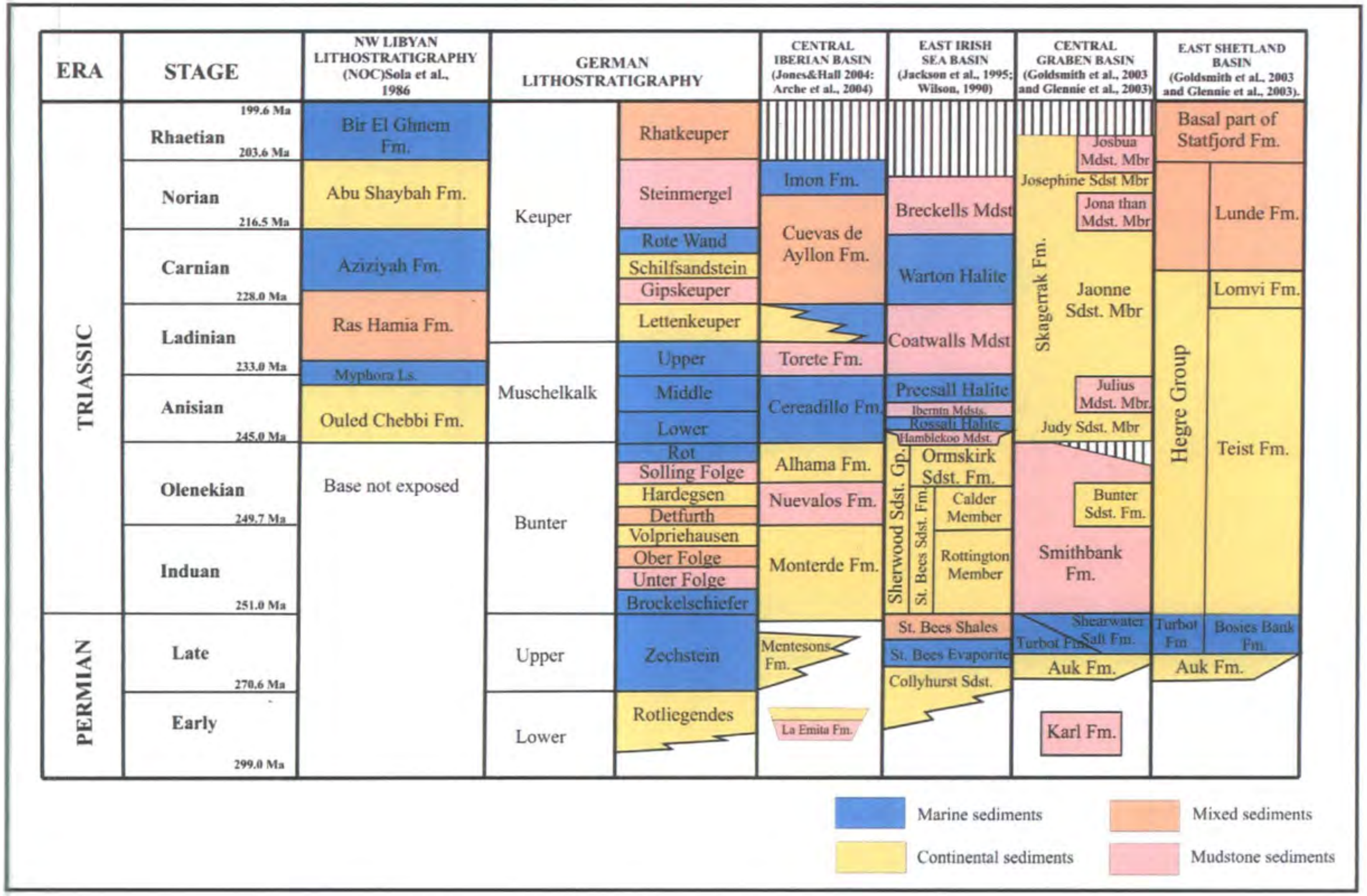


Fig. 1.2 Correlation panel for the Triassic Period in NW Libya and Europe (Adapted from Sbeta et al., 2005; Jones & Meadows 2007).

1.5 Triassic in North Africa

With the disassembly of Pangaea in the Early Mesozoic, the North African Platform became a broad Tethyan - facing passive continental margin (Stampfli *et al.*, 1991). A thick succession (>1.5 km) of Triassic to early Cretaceous sediments were deposited in a vast interior sag basin, the 'Triassic Basin of eastern Algeria, southern Tunisia and western Libya (Boote *et al.*, 1998).

During the Triassic, sediments were initially deposited in topographic lows, and progressively on lapped the Hercynian Unconformity as sea level rose (Carr, 2003). Rifting and crustal stretching propagated west-wards in Algeria, Tunisia and Libya during the Triassic with diffuse crustal extension (Guiraud, 1998). This was followed by more localised rifting in high, Middle and Saharan atlas and along the western margin of the platform during the Early Jurassic (Guiraud, 1998). Crustal separation followed in the Middle Jurassic with the opening of Alpine Tethys and the Central Atlantic (Guiraud, 1998).

The Triassic succession of southern Tunisia and NW Algeria, consists of a single recognisable non marine siliciclastic cycle of sedimentation grading up into shale and evaporites with carbonates associated with the Middle Triassic transgression. In south-eastern Algeria the Triassic succession consists of two identical depositional cycles each starts with alluvial fan deposits grading upwards and laterally into braided and meandering stream sandstones facies (Fig. 1.3).

The development of the supercontinent Pangaea allowed for deposition of Triassic red beds of great spatial extent, but due to the incipient rifting that was accelerating through the Triassic significant variations can be found in sedimentary facies and it is these variations that have important implications for understanding the geological evolution of the region and hydrocarbon plays. This section briefly discusses the spatial and temporal extent of the Triassic across North Africa with special emphasis to Libya.

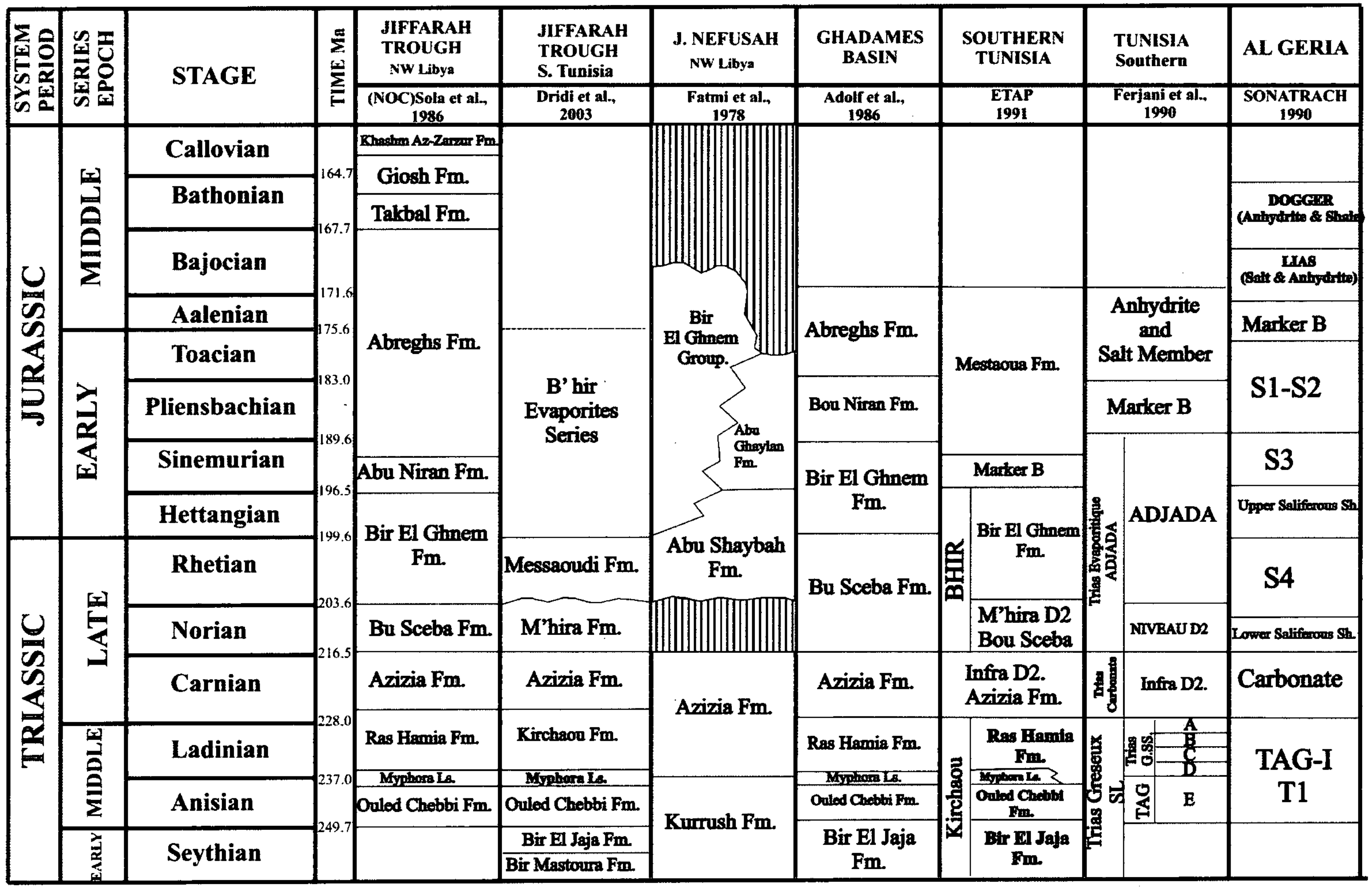


Fig.1.3. Stratigraphic charts to compare the Triassic successions in Libya, Tunisia and Algeria (Adapted from Sbeta *et al.,* 2005).

1.5.1. Triassic rocks in Morocco

During most of the Mesozoic in Morocco the High Atlas domain experienced extension and rifting, first during the Triassic, as recorded by red beds and tholeitic basalts, and later during the Jurassic, with the depositions of marine carbonates and shales capped by continental red beds. Though largely hidden below the Jurassic, the Triassic is thickest around the Massif Ancien, the Skoura culmination, and the southern Middle Atlas (up to 1000 m.). The reactivation of previous Hercynian faults has been considered responsible for the location of the Triassic basin, under a NW-SE tensional field (Pique *et al.,* 2000).

1.5.2. Triassic rocks in Algeria (Fig. 1.4)

In Algeria a blanket of Triassic fluvial sands and shales was deposited on the eroded Palaeozoic surface and these sands are overlain by extensive evaporates, which in turn are overlain by shelf carbonates. The Ghadamis Basin covers an area of approximately 250,000 km² in Algeria, Tunisia and Libya (Fig.1.4). The maximum thickness of sedimentary sections in the basin reaches up to 7000 m of mixed clastic and carbonate sedimentary rocks. The Algerian portion of the Ghadames Basin has recently been renamed the Berkine Basin and has been the focus for exploration activities and major new hydrocarbon discoveries. All of the high quality oil in the Berkine Basin is reservoired in Triassic sandstone plays. The Berkine Basin (Algeria) is an intra - or pericratonic basin that developed during the Middle to Late Triassic on the margin of the Saharan platform. The TAG-I sits unconformably on Palaeozoic basement rocks and with the basal Lower Carbonate comprises a laterally and vertically variable sequence which has been sub-divided into four depositional sequences: Sequence 1, an unconformity-bounded, ephemeral fluvial interval that fills inherited relief on the Hercynian unconformity; Sequence 2, an initially upward-fining, and subsequently upward-coarsening package of perennial fluvial sandstones and floodbasin shales with thin crevasse splay elements and interfluves' palaeosols; Sequence 3, an erosively based, fluvio-lacustrine section characterized by fluvial sandstones with associated crevasse sandstones and floodbasin/lacustrine shales. This sequence is the main hydrocarbon reservoir section and is divided into two main packages 3A and 3B; the base of 3B is distinguished by basin-wide fluvial incision and the widespread channel sand deposition; Sequence 4 is a coastal plain and shallow marine system comprising shales, sabkha-type evaporites and bay-fill sandstones (Turner et al., 2001).

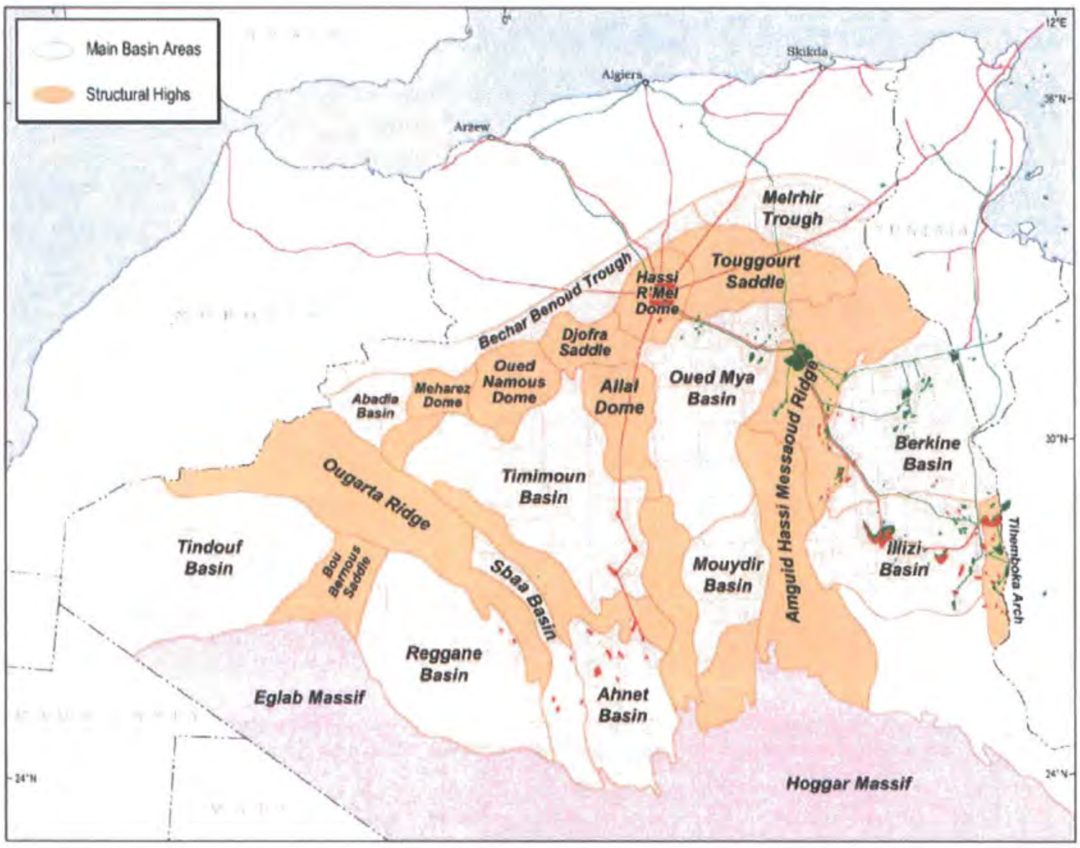


Fig. 1.4. Map showing the location of the Berkine Basin on the northern flank of the Saharan Platform, Algeria (From Turner *et al.,* 2001).

1.5.3. Triassic rocks in Tunisia (Fig. 1.5)

The Triassic deposits in southern Tunisia crop out widely in the Jifarah and are represented by sandstone with shale well bedded series. Northwards of the Saharan Platform the Triassic sediments are involved within fault zones and come through faults as diapirs or tectonic breccia. The Early Triassic rocks are up to 1600 m thick and consist of marine to fluvial sandstones and shale with local occurrences of limestone. Early

Triassic deposits occupy a narrow area and extend progressively and slightly westward and southward (Fig. 1.5). The Late Triassic rocks are up to 900 m thick and consist mainly of two lithological intervals, from base to top: a siliciclastic interval overlain by mixed carbonates, siliciclastics and evaporites. The Late Triassic deposits occur virtually throughout a very extensive area. The deposits began in downwarp basins in the Jifarah area and extended progressively onto the Saharan Platform. The high areas were overlain, except for some persistent apexes such as the Tebaga of Medenian, by a Late Triassic series. A complete Triassic stratigraphic section is known in southern Tunisia and crops out along a NW-SE trending belt from Jabal Tebaga of Medenine to the Libyan border and continues eastward in Libya along Jabal Nafusah. In the Jifarah Basin, the succession is thick and consists of Early to late Triassic sediments, usually lying unconformably on the Permian or older Palaeozoic rocks or basement. On the Saharan Platform, the succession is thinner and consists of Late Triassic sediments overlying older Palaeozoic strata or basement. Northwards of the E-W Chotts trend, the Triassic series crop out by injection through faults and within diapirs. In the vicinity of Gharyan, mainly Late Triassic series crop out.

The Triassic sandstones constitute the major exploration target and the main producing reservoirs in southern Tunisia. They provide the main reserves of hydrocarbon in the Ghadamis Basin (Tunisia side). El Borma oil field in southwestern Tunisia is located on the Algerian border and produces from five Upper Triassic sandstone reservoirs at depths ranging from 2,300 to 2,400 m.

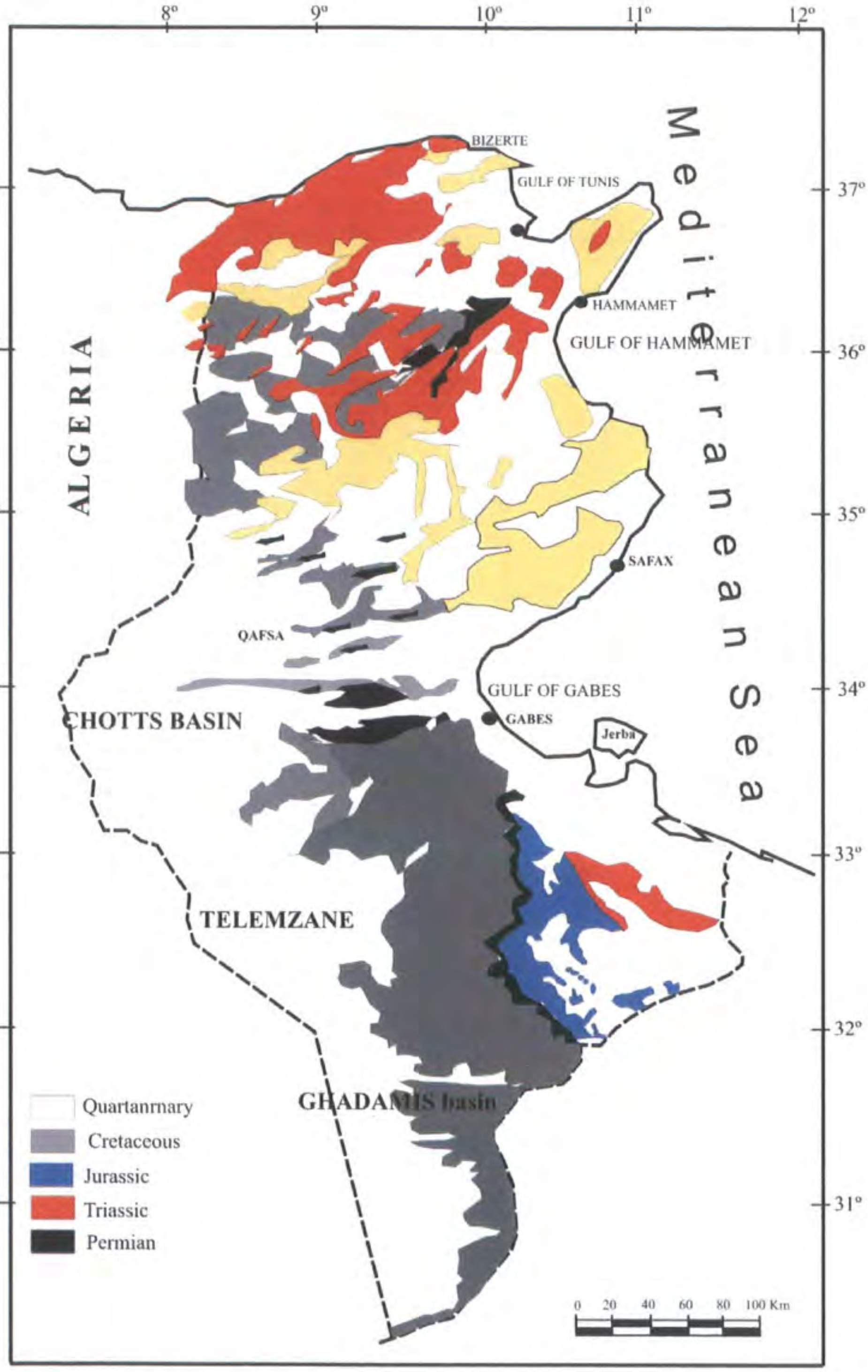


Fig. 1.5.Simplified geological map of Tunisia showing the distribution of Triassic rocks in northern and southern Tunisia (Dridi and Maazaoui 2003)

1.5.4. Triassic rocks in Libya (Fig. 1.6)

Mesozoic rocks are present over much of Libya, with the exception of the principal Hercynian uplifts. Rifting began in the Triassic and a series of horsts and graben developed on the Sirt Arch, some of which contain Triassic and Jurassic sediments. During this period marine rocks were confined to the northern margin of Libya and the interior was dominated by continental deposition (Hallet. 2002) (Fig. 1.6). Triassic rocks in Libya occur in four main domains: (i) Predominantly marine rocks of the Nafusah escarpment and offshore, (ii) The eastern extension of the Algerian Ghadamis area, (iii) Subsurface Triassic in eastern Libya, and (iv) Continental rocks of the interior. Rubino et al., 2002, divided the Triassic sequence of the Nafusah escarpment into four sequences separated by unconformities. Sequence 1 equates to the Al Guidr / Kurrush Formations, Sequence 2 to the Al Aziziyah, sequence 3 to the Abu Shaybah Formation and Sequence 4 to the Abu Ghaylan carbonates. In the offshore the Triassic is mostly too deep to have been penetrated by wells but seismic data shows thick Triassic sediments to the north of the Sabratah-Cyrenaica Fault. Triassic rocks are well developed in the Sabratah Basin and evaporites are present in the late Triassic in the western part of the basin. Over 1000 m of Triassic sediments and the sequence thickens gradually north wards, reaching 1500 m in the Misraratah Basin and 3000 m on the Medina Bank. In the Sirt Basin the Triassic is thinner ridge not more than 500 m., and less than 200 m on the Cyrenaica Ridge (Fig. 1.6).

Recent years major new information has been obtained from wells in the north eastern Sirt Basin which suggests that the rifting phase in the Sirt Basin began as early as the Triassic. A palynological study of pre-Upper Cretaceous section in well A1 -96 (Jakharrah field), a unit formerly equated with the Amal Formation, yielded a rich palynomorph assemblage of late Scythian to Middle Triassic age. This unit consists of organic – rich shales and sandstones with some oil-source potential, deposited in conditions alternating between oxic and anoxic. The thickness reaches 440 m at Jakharrah, Sirt basin, (Thusu and Vigran, 1985).

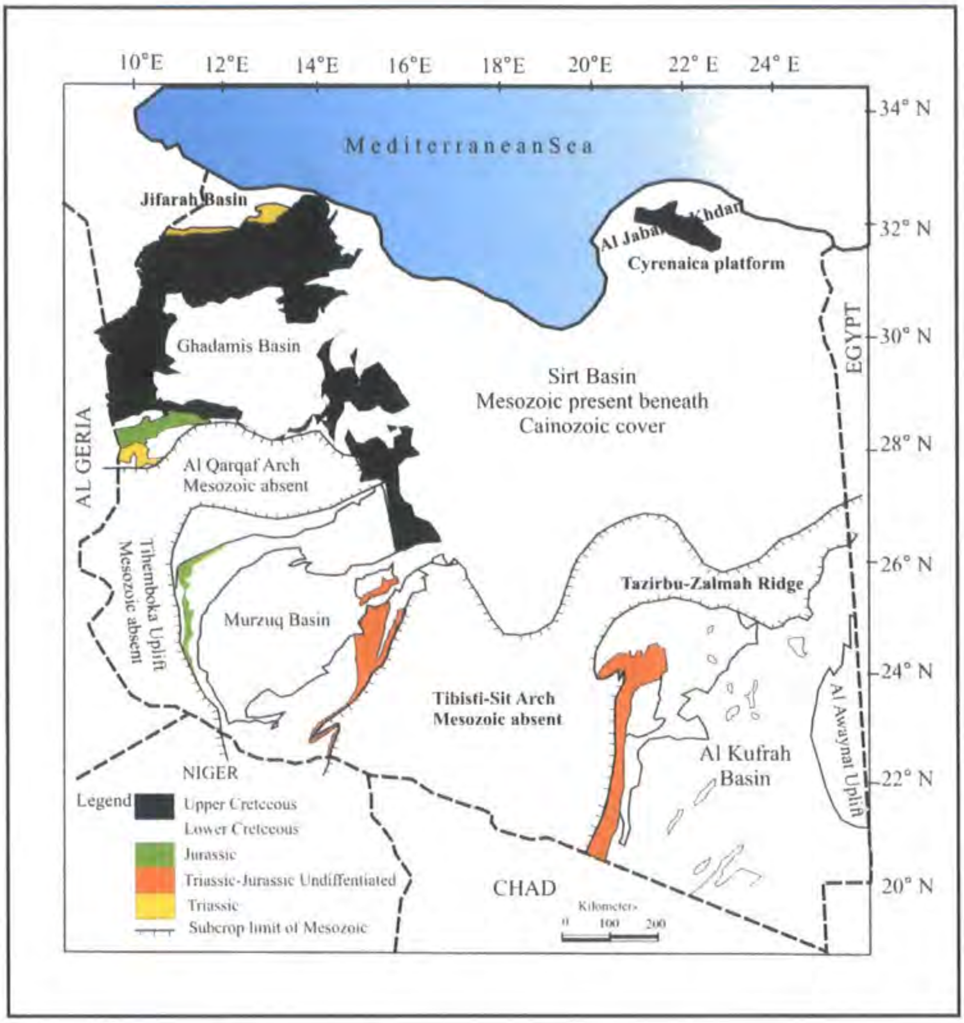


Fig. 1.6. Mesozoic Outcrops and Subcrop Limit, Sources: Map of Libya; (Scale 1: 1000, 000, Wennekers *et al.*, 1996).

1.5.4.1. Jabal Nafusah and Jifarah basin (Fig. 1.7)

The generally flat and fertile though somewhat undulating coastal plain, known as the Jifarah Basin, rises towards the south into an impressive escarpment forming hill range known as Jabal Nafusah. The trend of the Jabal is proximately ENE-WSW extending from near Mesellata in the east where it is closer to the Mediterranean coast to Nalut and beyond the international border into Tunisia in the west (Fig. 1.7). The central part of the Jabel is located approximately 70 Km south of Tripoli. The escarpment is irregularly distributed in its east-west extent due to dissection by ravines and wadies cut in the Mesozoic rocks. To the south of the escarpment lies the broad dissected plateau Al

Hammadah al Hamra extending towards Mizdah and Ghadamis exposing mainly Upper Cretaceous rocks with some Paleocene cover.

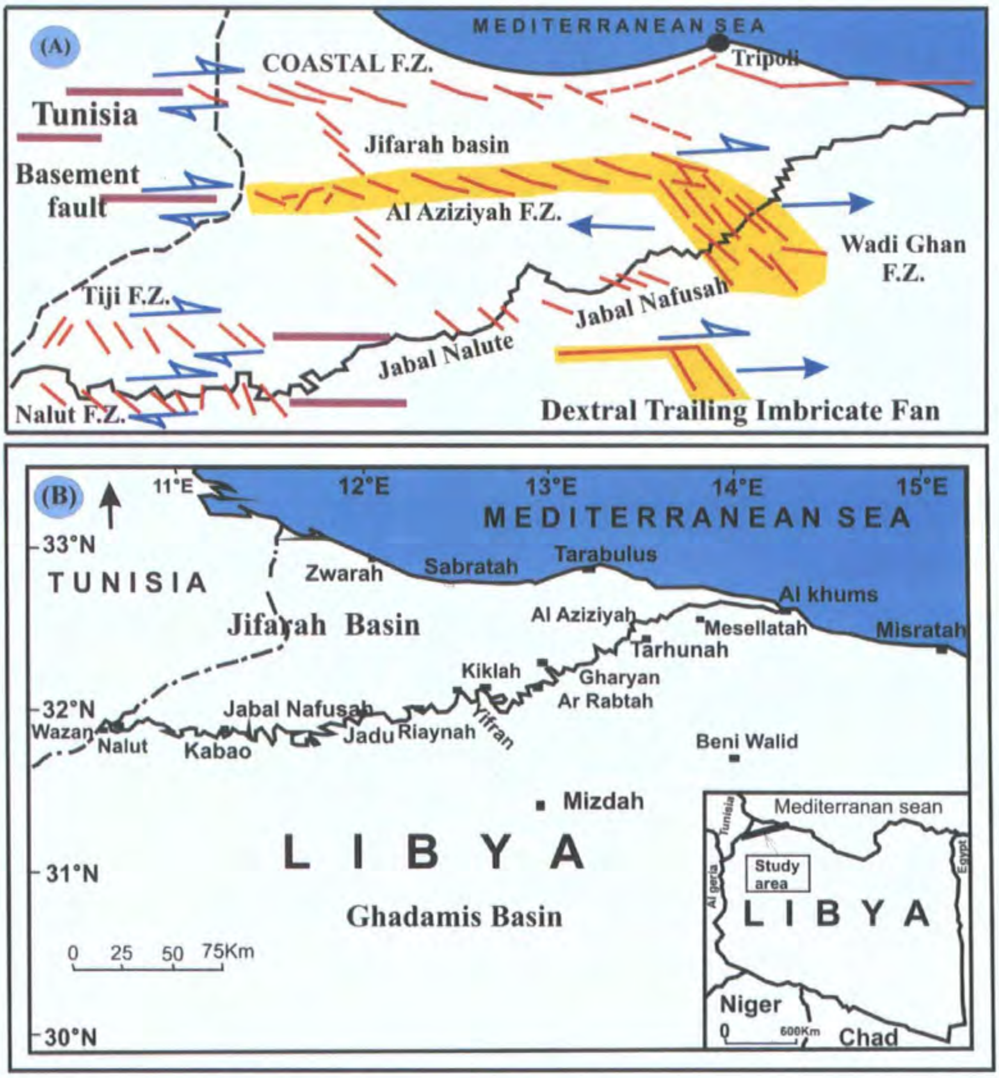


Fig. 1.7. Location maps of the studied area, Map (A); Interpretation of Al Aziziyah fault zone (Al Aziziyah F.Z.) and Wadi Ghan fault (Wadi Ghan F.Z.) in terms of dextral strike-slip faulting in studied area. Such a relation was used as a model to interpret the relationship between South Cyrenaica fault zone and Sirt basin rift (Source: Anketell and Ghellali, 1991; Anketell, 1996). Map (B) General location map of Jifarah Basin and surrounding area.

In the eastern part of the Jabel (Khums and Messellata area) the Cretaceous sequence is overlain by the Miocene rocks. The escarpment is capped by hard Upper Cretaceous (Cenomanian –Turonian) formations (Sidi As Sid and Nalut Formations) which stand out prominently forming steep slopes and cliffs. The lower parts of the escarpment face towards the north exposing older Pre Upper Cretaceous rocks. The sedimentary sequence of Jabal Nefusah is mainly of Mesozoic in age but in no single section the sequence is complete and is interrupted by important unconformities some of which are well established while others are disputed.

The Mesozoic sedimentary sequence shows a considerable variation in lithology and thickness which reflects the shifting of sedimentary environments both laterally and vertically. The Early Triassic rocks are poorly exposed (Lower member of Kurrush Formation) and indicate a continental environment with the development of red micaceous sandstone and siltstone facies. The rocks of early Middle Triassic age show a mixed carbonate and terrigenous lithology formed in an alternating near shore marine to a continental environment. In the latter part of Middle Triassic shallow water marine conditions were well established in the area which resulted in the deposition of limestone and its micro facies (lower member of Al Aziziyah Formation).

During the early Late Triassic (Carnian) there was some influx of terrigenous material in the basin (Upper mixed carbonate and terrigenous material of the Al Aziziyah Formation) which was followed by an emergence at the close of Carnian (Unconformity between Al Aziziyah and Abu Shaybah Formations). The overlying Abu Shaybah of doubtful Late Triassic to Early Jurassic age was deposited in a continental environment (fluvial) that gradually shifted to a shallow water marine (Abu Ghaylan Formation) in the central Jabal to Evaporate (near shore lagoonal) environments in the Western Jabal (Bir al Ghanam Group) in the latter part of the Early Jurassic (Fig. 1.7).

The Abu Shaybah Formation (ASF) is exposed from the foothill slopes of the Tarhuna - Gharyan scarp stretching westwards to Ar Rabitah and Al khums to the eastward along the Jabal Nafusah. The Lower boundary is sharp and unconformable with the Al Aziziyah Formation, (marine deposit) and the upper is locally unconformable with the Abu Ghaylan Formation (marine deposit).

Seven sections have been measured in the study area and the maximum thickness of ASF is about 254 m located in Wadi Ghan. This is the best-exposed outcrop on Jabal Nefusah to understand the ASF as a whole and the contacts with both underlying Al Aziziyah and overlying Abu Ghaylan Formations. Although the ASF is quite thick (>250 m), the topmost unit present along the main road to Gharyan, which displays lagoonal facies above the fluvial, is truncated by the pre-Abu Ghaylan unconformity.

In recent years important new discoveries have been obtained from wells in the north-eastern Sirte Basin which suggests that the rifting phase in the Sirte Basin began as early as the Triassic. A palynological study of the pre-Upper Cretaceous sections yielded a rich palynomorphic assemblage of late Scythian to Middle Triassic age. Such new findings have heightened interest in the Triassic of the Jifarah Basin as an important field analogue with excellent exposures of Late Triassic red beds that possibly can be found subsurface in the Sirte and Al Kufrah Basins.

1.6. Summary

The wide extent of Triassic across North Africa has long been recognised as an important area for hydrocarbon exploration and over the last decade most exploration has concentrated on the Triassic of the Ghadames Basin, Algeria. In the Ghadames Basin at least 20% of production comes from the Triassic with Algeria, Libya and Tunisia conducting exploration activities in adjoining parts of the same basin, all concentrating on Triassic reservoirs.

This chapter has highlighted the importance of research into the Triassic of North Africa but also the similarities with European sections (e.g. Central Graben, North Sea, Central Iberia, Spain, and Germany) where better temporal frameworks have already been established.

The Triassic of the Jifarah Basin of NW Libya allows for one of the first detailed studies of a Late Triassic fluvial system isolated from the shallow marine facies that dominate the North African rift basins to the west. The fluvial succession of the Abu Shaybah Formation shall be evaluated in the following chapters in terms of the main controls on facies distribution, petrography and reservoir quality in low-net to gross systems. This will be used to establish an improved model of the fluvial system within a sequence stratigraphic framework.

Chapter 2
Structural and Tectonic Framework of Libya

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2.1. Tectonic Framework

The early regional geological mapping of Libya, principally by Italian geologists, identified four major sedimentary basins with hydrocarbon-bearing potential: Sirt Basin, Ghadames Basin, Murzuq Basin and the Al Kufrah Basin (Fig. 2.1). The Sirt Basin is a major sedimentary basin that extends southwards from the Gulf of Sirt in central Libya. Unlike the basins of southern Libya there is little surface expression of the Sirt basin. It lies beneath vast sarir gravel plains, with occasional sand seas and escarpments. The existence of the Sirt basin was unknown until gravity and magmetic surveys were carried out in the late nineteen-fifties as part of the quest for petroleum. Subsequent seismic surveys and drilling have proven a major petroleum province, with estimated reserves in excess of 45,000 million barrels of oil and gas oil equivalent (Thomas, 1995). Despite over 35 years of petroleum exploration, however, relatively little has been published about the Sirt basin. The Ghadames Basin and the Murzuq Basin are the areas of primary interest in western Libya. These two basins principally contain Palaeozoic hydrocarbon-bearing strata and correlation to the other Libyan basins. The Sirt Basin occupies terrain in the east and northeast of the country and hydrocarbons are mainly contained in Mesozoic and Cenozoic sediments. In comparison Al Kufrah Basin has seen very little hydrocarbon exploration and is relatively unknown. In addition to onshore tectonic basins of Libya, signatural exploration has been undertaken offshore in the Mediterranean Sea, where most work has concentrated to date on the Cenozoic with the basin formation probably dating from the Carboniferous.

Libya is presently situated on the northern margin of the African continent and, its Palaeozoic basins are a product of the plate tectonic and geological events associated with the early evolution of the African continent as part of the ancient Gondwana and Pangaea supercontinent. In contrast, the Mesozoic and Cenozoic basins evolved in a predominantly tensional, wrench tectonic setting controlled by the relative motion of the

African and Eurasian plates in the Tethyan and then Mediterranean domain (Hallett. 2002).

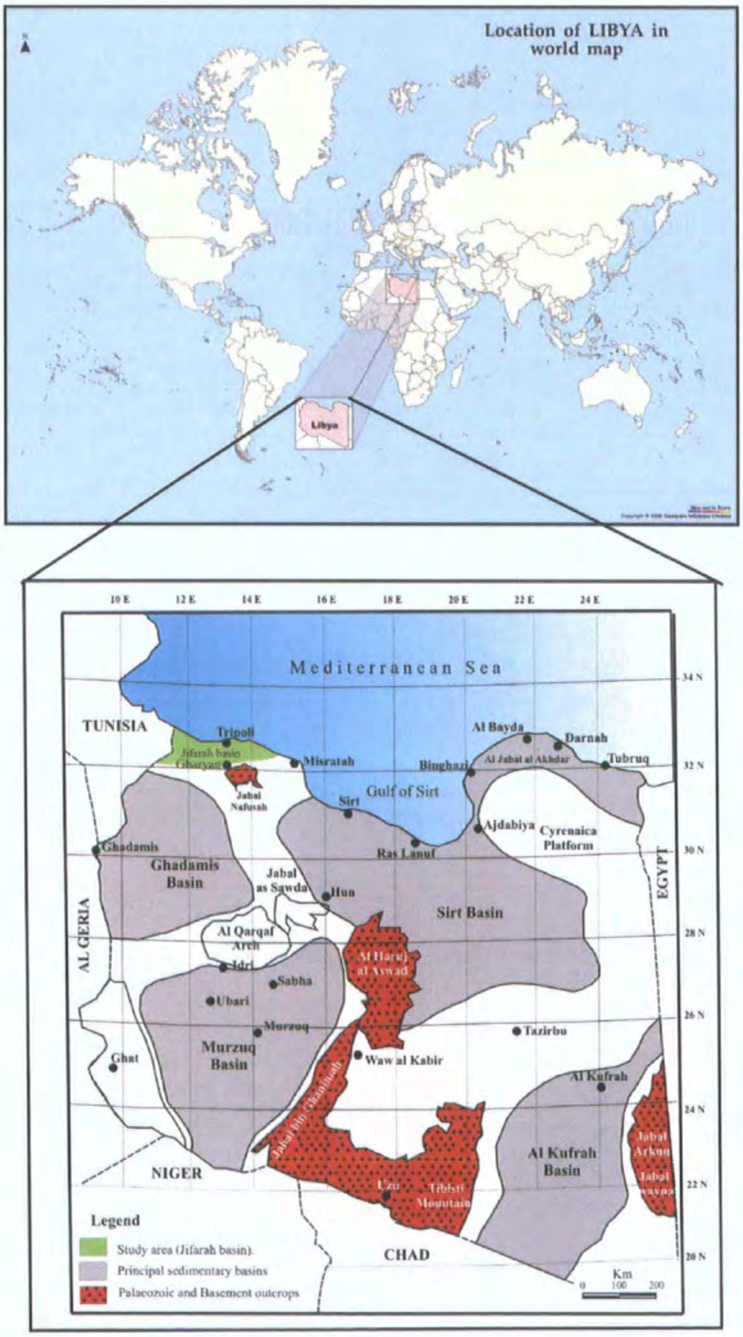


Fig. 2.1. Showing Sedimentary basins of Libya and studied area of Jifarah basin (Adapted from Geological Map of Libya, Futyan and Jawzi, 1996).

This chapter begins with a more detailed outline of the Palaeozoic and Mesozoic basins of Libya, which have a considerable bearing on the understanding of the Triassic in the Jifarah basin of NW Libya (section 2.2). The next section continues with a description of the Jifarah region that forms the focus of this study (section 2.3). The stratigraphy and depositional model for the Triassic of Northern Libya is given in section 2.4 and 2.5. The final section provides a summary of the key aspects of development of the Triassic and especially in the Jifarah Basin (section 2.6).

2.2 Sedimentary basins of Libya

2.2.1 Sirt basin (Fig. 2.2)

The history of the Sirt Basin (Fig. 2.2) started in the Triassic, when early rifting preceded the first formation of Jurassic oceanic crust in the Eastern Tethys (Hallett, 2002). The Eastern Tethys was the northern extension of the Central Atlantic (Western Tethys) spreading axis, connected to the latter by a "leaking transform" zone south of Iberia. The sinistral shear zone was caused by the blocking, in the region of what is now the Austrian Alps, of a north-eastern African complex of continental blocks, which could not follow the overall eastward movement of the African continent as a whole relative to Europe.

In the Sirt Basin, Triassic rift sediments are rare, and where present, relatively thin and not associated with the main grabens. It could be argued that the Triassic sediments of the Sirt Basin are not related to Eastern Tethys basin development, but rather to an overall sinistral shear tectonic setting in a late sag phase of the old Tethys margin. The marine-influenced lagoonal facies of the Middle and Late Triassic in the extreme northeast of Libya and on the northern Cyrenaica Platform would support this, as marine influence can only have come from the "Alpine Triassic" to the northeast.

The undifferentiated Triassic of the present-day western Sirt Basin can be seen as an intramontane continental facies developed in a depression on the Libyan Arc (Hallett, 2002). It seems significant that this occurrence is, for the major part, on the Waddan Uplift and not associated with any Sirt Basin troughs, because these postdate the Triassic depression. Similarly, non-marine Triassic occurrences on the Rakb Field, (Abadi, 2002),

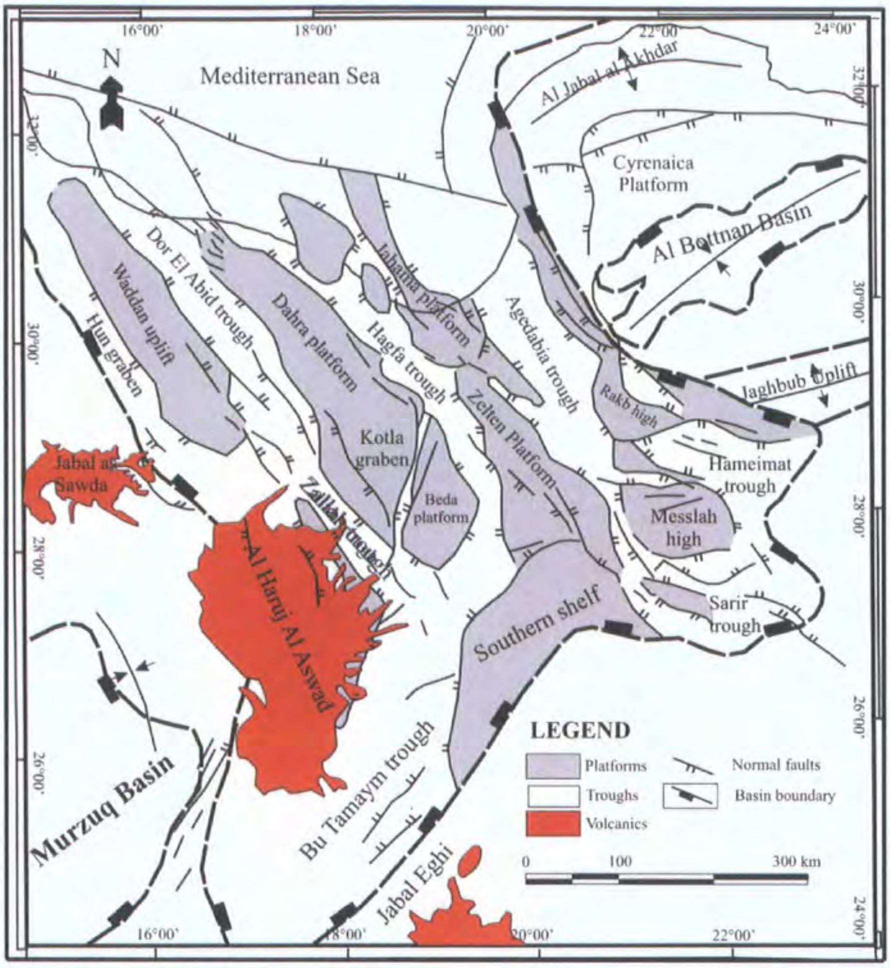


Fig. 2.2. Location map of Sirt Basin. Troughs and grabens, platforms and horsts are synonymous terms. (Adapted from Abadi, 2002).

are on a structural high rather than in a trough. Abadi (2002) did not include a Triassic phase in his multi-phase rifting history of the Sirt Basin, thus implying that the Triassic was pre-rift. Whilst this may be correct relative to the later Jurassic rift system, Hallett (2002) is probably also right in arguing that there must have been Permo-Triassic rifting causing magmatism (Wilson and Guiraud, 1998). The discrepancy in interpretation lies in Hallett's (2002) linking Triassic rifting to the development of the Eastern Tethys and the formation of the Sirt Basin, whereas Abadi (2002) considers it to be independent of that. Hallett (2002) states that a main rifting phase occurred in the Middle Jurassic. However, there is little evidence of Middle Jurassic grabens in the Sirt Basin. Marine Jurassic rocks are confined to the extreme northwest of Libya and to Cyrenaica, with WSW-ENE trending facies belts and general thickening towards the NNW. According to Hallett (2002), Jurassic rocks are also present below the Sirt embayment, extending into the

Ionian abyssal plain. Obviously, these connect the northwest Libya and Cyrenaica Jurassic. Undifferentiated Jurassic sediments in Nubian Sandstone facies are known from some western and central parts of the onshore Sirt Basin, but these are probably not older than Tithonian and related to a later rift phase. Jurassic basin configuration is broadly parallel to the general direction of the Eastern Tethys rift system, which, however, had reached the spreading phase by Bathonian/Callovian times. This is almost exactly the time when marine Jurassic deposition in northwest Libya started (Takbal Formation, Early Bathonian). Therefore, the marine Middle and Upper Jurassic can perhaps better be explained as Eastern Tethys post-rift, unless there is independent evidence of the rift origin of the Sirt Basin Jurassic from seismic data (e.g. oldest sediments in grabens and graben fills with diverging reflector patterns).

NW-SE and E-W trending grabens with rift sediments developed from the latest Jurassic onward. The syn-rift graben fills are in Nubian Sandstone facies, which sometimes show a tripartite sandstone-shale-sandstone subdivision. The Nubian Sandstone reaches a maximum thickness of 1200 m in the Hameimat Trough. The shales were dated as Berriasian to Barremian and may well contain a rift sequence maximum flooding surface which coincided with the acme in rifting activity in the Eastern Mediterranean. The structural grain of the Sirt Basin therefore relates to the Late Jurassic/Early Cretaceous rifting and opening of the Eastern Mediterranean rather than to that of the Eastern Tethys *sensu* Hallett (2002), in which one would expect NE-SW trending horsts and grabens.

The opening of the Eastern Mediterranean in the Middle Cretaceous (Aptian/Albian) is probably related to the rotation of the "Apulian" or "Adria" plate, i.e. the complex of continental fragments which were coupled with Europe in the Eastern Alpine region and therefore could not follow the south-eastward movement and more gentle counter clockwise rotation of the African continent Hallett's (2002). 150 Ma and 120 Ma plate tectonic reconstructions (after Dercourt *et al.*, 1986) illustrate that there is no real other reason why the Eastern Mediterranean, could have formed. Tectonic subsidence analysis by Abadi (2002) showed that a relatively rapid Cretaceous subsidence phase ended ca. 112 Ma ago; this is about Aptian, when spreading in the Eastern Mediterranean probably started. Nubian sandstone deposition (Upper Cretaceous) temporarily came to an end and there was erosion in many places, suggesting tectonic adjustment due to the ending of the tensional regime. The ensuing unconformity is a classic break-up unconformity; it separates the Upper Nubian (Upper Cretaceous) from the older units.

Widespread subsidence started again in the Cenomanian, considering unit a marine transgression. All Late Cretaceous sediments were deposited in broad sag superimposed on the NW-SE and E-W trending platforms and troughs, with overall on lapping, transgressive relations to older rocks in the Sirt Basin. Due to the relative movements of the European and African continental blocks, oceanic crust was gradually subducted to the North. This triggered the formation of back-arc basins, the most mature of which is the Algero-Provençal/Alboran Basin, which thrust the Kabylian continental fragments onto the North African margin and moved Sardinia and Corsica from the southeastern margin of Iberia and southeast France into their present-day positions.

The Tyrrhenian Sea was still spreading, pushing Calabria east-southeastwards, with an accretionary wedge ahead of it and a volcanic arc behind. Northward subduction also created the Aegean back-arc basin in the Eastern Mediterranean, but this has not yet created oceanic crust, only a double volcanic arc. During all this tectonic activity, the Sirt Basin was on the continental part of the African plate. The Cenozoic tectonic history of the Sirt Basin is marked by rather subtle changes in stress regimes due to tectonic settings and events in a broader regional context. The Sirt Basin was always quite remote from any plate boundaries and thrust fronts. The nearest major tectonic feature is the frontal thrust of the Aegean arc, which is more than 450 km away from the onshore Sirt Basin. The regional tectonic setting has been summarised adequately by Hallett (2002). For the basin fill, this means that there are no major inversions of troughs and drowning of individual platforms. All major structures remained as they were at the end of the Early Cretaceous, with adjustments due to regional, not local, stress regimes. An exception must be made for Oligocene and Neogene volcanism which was widespread along the Tripoli-Tibisti axis, to the west and south of the Sirt Basin. It is not entirely clear what this volcanism is related to, perhaps the passing of the Cameroon hotspot (c. 14 Ma), followed by the Hoggar hotspot. These also caused relative uplift in western areas, where

Cretaceous and Tertiary Sirt Basin sediments are now in outcrop, whereas sedimentation largely continued well into the Neogene and the Quaternary further east.

The above discussion assumes that the last real rifting, in the sense that there was a nearby rift basin of which the Sirt Basin can be seen as a failed branch, was Aptian-Albian, i.e. the rifting of the eastern Mediterranean basin. All post-Albian (? Aptian) is post-rift in that sense, because the eastern Mediterranean literally broke up and was spreading. Later irregularities in basin subsidence were caused by changes in stress fields in response to documented changes in the plate motions of Africa relative to Europe, the Atlantic and the Tethys/Mediterranean area as well as rifting in other African Basin, but not by the formations of new rift basins. According to Abadi (2002) the subsidence analysis shows that the Sirt basin is characterised by a number of pulsating phases of rapid tectonic subsidence followed by episodes of slower rates. He identifies a Palaeozoic pre-rift stage, three Late Jurassic to Early Eocene syn-rift stages and a post-rift basin fill of Neogene strata based on the back stripping analysis of 225 wells in the Sirt Basin.

2.2.2 Jifarah basin (Fig. 2.1)

The Jifarah basin is bounded by the Jifarah Fault and Nafusah Uplift to the south and the Sabratah-Cyrenaica fault system and Sabratah Basin to the north, and represents a downfaulted terrace on the unstable continental margin. It underlies the Jifarah Plain and extends westwards into Tunisia north of the Dahar Uplift covering an area of 15,000 Km^2, more than half of which is in Tunisia.

The basin has been penetrated by a numbers of wells (mostly water wells) which reveal a stratigraphy extending from Palaeozoic to Jurassic in age, overlain by a thick Miocene cover. Permian rocks, encountered on the northern flank of the Nafusah Arch in wells in Libya, and Tunisia, are presumed to continue northwards, forming basement in both the Jifarah and Sabratah Basin. North of the Jifarah Fault a thick Permian sequence has been found in wells in Tunisia and Libya. This sequence ends abruptly at the Jifarah Fault. The Jifarah Fault is part of the Sabratah-Cyrenaica wrench zone, which marks the boundary to the relatively stable shelf to the north. The Mesozoic basin –fill of the Jifarah Basin was affected by syn-depositional faulting during the Triassic and by shearing during the Neocomian which reflect major tectonic activity along the southern margin of Tethys. Anketell and Ghellali (1991) demonstrated that the dominant fault directions in the Mesozoic section are east-west and NNW-ESE with an en echelon arrangement which they interpreted as riedel shears and imbricate fan splays formed as a result of strike-slip movement on the South Atlas-Jifarah dislocation (Fig. 2.3).

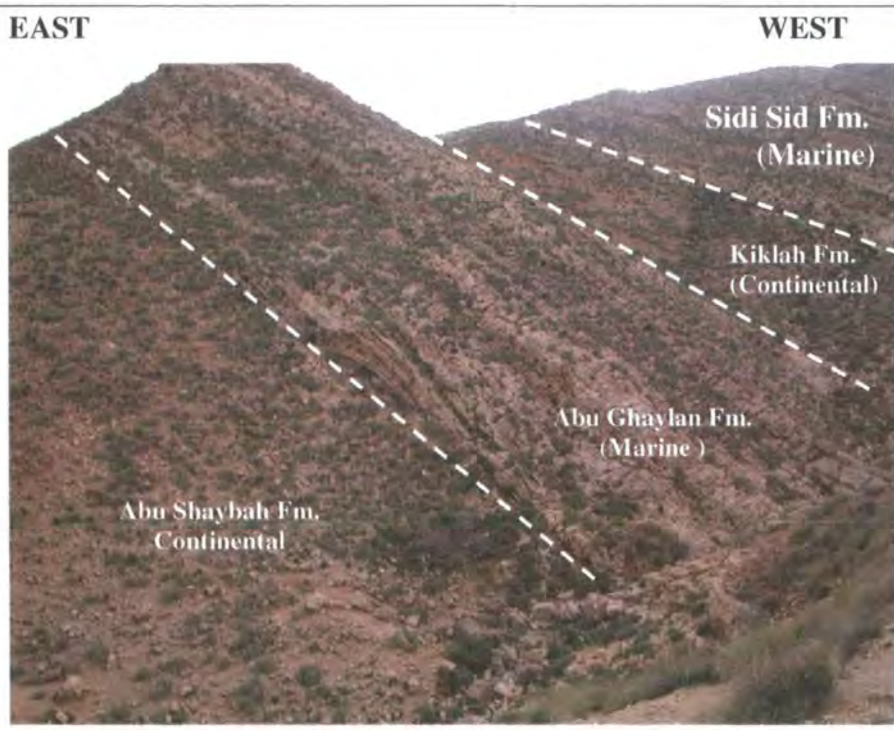


Fig. 2.3 Showing Abu Shaybah – Abu Ghaylan Formation and fault East – West direction, Abu Shaybah Late Triassic, Abu Ghaylan Early Jurassic, Kiklah Fm. Early Cretaceous (Aptian-Albian) and Sidi Sid Late Cretaceous (Cenomanian).

They developed a model which suggested an analogy between the imbricate fan splays visible in the Wadi Ghan area south of Al Aziziyah with the strikingly similar fault trends in the Sirt Basin, and attributed both to strike-slip faulting associated with underlying basement dislocations.

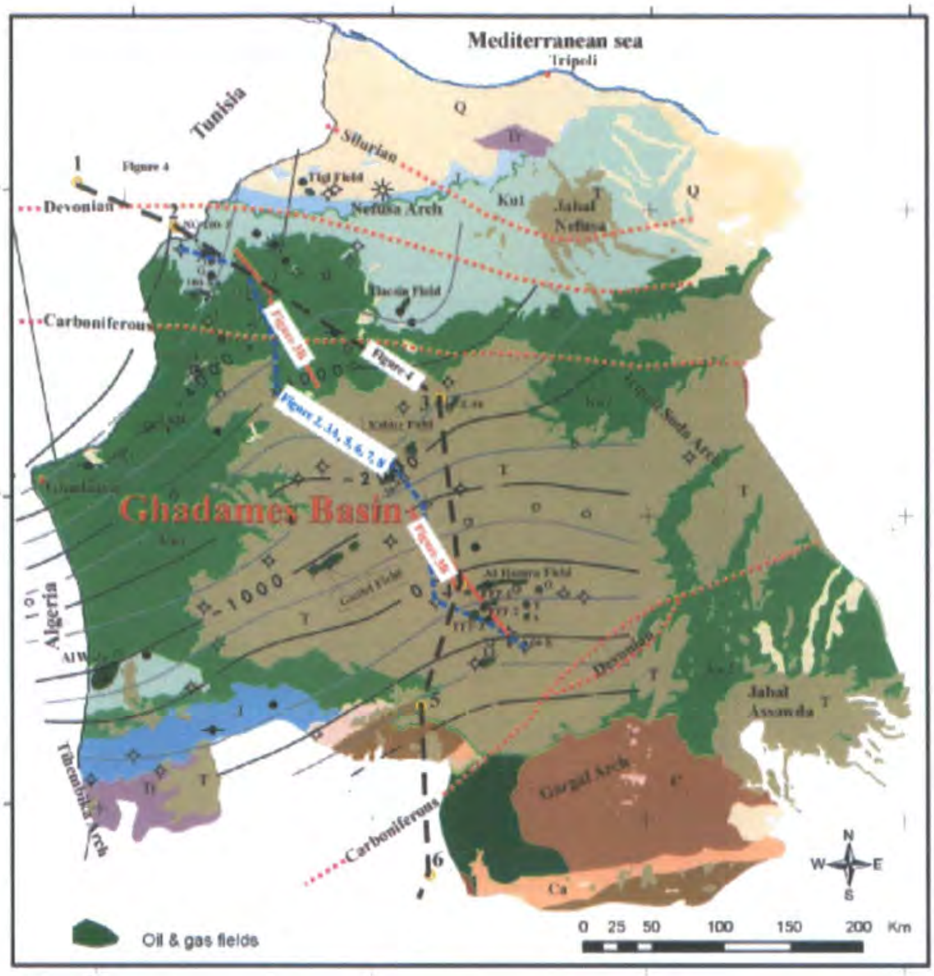
Sedimentation continued until the Palaeocene, but the basin was caught up in the Eocene tectonism which reactivated the Nafusah Arch and introduced a new generation of faulting in the Jifarah Basin. Extensive pre-Miocene erosion removed much of the Mesozoic sediments from the basin, and the pre-Miocene subcrop shows a complex pattern. Wells show Miocene rocks resting on Lower Jurassic Bir al Ghanam Formation, whereas water wells to the west show Miocene rocks resting on Middle Triassic Kurrush Formation. By comparison, on the Nafusah escarpment south of the Jifarah Fault, almost the entire early Mesozoic sequence is preserved since the beginning of the Miocene the basin has been subsiding towards the north.

2.2.3 Ghadames basin (Fig. 2.4)

The Ghadames Basin (Fig. 2.4), like the Al Kufrah (2.2.1) and Murzuq Basins (2.2.4), is a large intracratonic sag basin developed on the passive northern margin of Gondwana. It covers an area of 350,000 km^2, with the basin centre located in Algeria. The Libyan portion represents the eastern flank of the basin, rising towards the Tripoli-Tibisti Uplift, with a small sub-basin, the Zamzam Depression extending towards Misratah. The basin contains up to 6000 m^2 of the basin fill in Algeria, but not more than 5200 m in Libya. The basin is bounded in the west by the Amguid -El Biod Uplift in Algeria, to the south by the Hoggar Massif in Algeria and the Gargaf Arch in Libya, and to the north by the Dahar-Nafusah Uplift. To the east the basin wedges out beneath the western part of the Sirt Basin.

The most conspicuous feature of the basin is the Hercynian unconformity which truncates the Palaeozoic succession and is overlain by a Mesozoic basin (the Hamadah Basin in Libya) with a markedly different basin configuration to that of the Palaeozoic. The tectonic history of the basin has some similarities with the Murzuq Basin (Hallett, 2002).

The Palaeozoic history of the basin was controlled by the northwest-southeast Pan-African tectonic trend, although little evidence of this has come to light due to the thickness of the overlying sediments.



Key: C=Cambrian, O=Ordovician, Ca=Carboniferous, Tr=Triassic, J= Jurassic, Ku1=Cenomanian-Turonian, Ku2=Coniacian-Maastrichtian, T=Tertiary, Q=Quaternary

Fig. 2.4. Location map of Ghadames Basin, (From Dardour et al., 2004).

The basin narrows southwards, confined between the Tripoli-Tibisti and Tihamboka Uplift, into the Murzuq Basin. To the east the Palaeozoic section pinches-out against the Tripoli-Tibisti Uplift, but to the west the Tihemboka Uplift did not extend further north, and the basin widened into a broad depocentre extending into Tunisia and Algeria west of the Tihemboka Uplift. Basement in much of this area is formed by Pharusian accreted terranes, but further south, and particularly in the Illizi Basin of Algeria, it is represented by rocks of the Pan-African remobilized belt.

The final pulses of Pan African tectonism continued into the Ordovician. During Landeilian times, uplift and erosion occurred on the Tihemboka Arch and the Ahara Uplift in Algeria, and during the Caradocian, folding, faulting, uplift and erosion occurred which removed much of the early and middle Ordovician section on the Dahar Uplift in Tunisia. It was onto this irregular surface that the Late Ordovician ice-sheet encroached from the south. Volcanic rocks were produced during these final tectonic phases which have been penetrated in several wells in the eastern Illizi Basin.

Tannezuft shales (Silurian) were deposited throughout the basin, but Rhuddanian radioactive shales are confined to the early depocentre where euxinic conditions existed prior to the basin wide flooding event (Hallett 2002). Early Devonian tectonic activity led to uplift and erosion of the Akakus Formation over large parts the basin. This is evident from well evidence in the Illizi Basin where only the Lower Akakus Sandstone is preserved, and similar truncation has been recorded in Libya. Folding of this age is evident on the Kabir trend in Libya where early Devonian sandstones unconformably overlie folded Akakus sands. Over most of the Ghadames Basin rocks of Tournaisian age are absent and in the Illizi Basin there is evidence of thickness variations and erosion on the northwest flank of the Tihemboka Uplift which indicate continued tectonic activity during the early Carboniferous (Hallett, 2002). The Hercynian orogeny reached its peak during the Late Carboniferous and major new tectonic elements were formed, including the Qargaf and Nafusah Uplifts in Libya, the Dahar Arch in Tunisia and the Talemzane and El Biod Arches in Algeria. The entire area was uplifted and subjected to intense erosion during the Permian which left the basin surrounded by highs which, in the case of the Nafusah and Qarqaf Arches, were eroded to their Cambro-Ordovician roots. A subcrop pattern of progressively younger rocks can be traced into the centre of the basin where a complete section up to Late Carboniferous is preserved. Within the Libyan part of the basin structural trends with a northeast- southwest alignment were formed during the Hercynian upheaval (Hallett, 2002).

Thick successions of Mesozoic continental deposits, including Triassic sandstones and evaporites, were deposited in the post Hercynian sag basin (Known as the Triassic Basin in Algeria and the Hamadah Basin in Libya) in which the depocentre was located much further north than during the Palaeozoic. The basin was affected by an extensional episode during the early Jurassic which produced block faulting in eastern Algeria. Jurassic transgressive sequences are followed by an early Cretaceous regression which was terminated by deformation during the Aptian related to the opening of the Mediterranean Sea spreading axis in the southern Tethys. Wrenching and uplift of Aptian age has been reported from the Illizi Basin and the Tihemboka Uplift was reactivated, faulted and uplifted.

More emphasis is required in the Triassic of the Ghadames Basin and so lay as only brief the TAG -I after it is the Triassic that is important.

2.2.4 Murzuq basin (Fig. 2.5)

The Murzuq basin (Fig. 2.5) covers an area from Algeria to Libya and in southwestern Libya is occupied by a classic sag basin, whose sub circular shape is clearly visible on satellite photographs. It covers some 40,000 Km^2, extending southwards into Niger. The Murzuq basin is separated from the Illizi basin to the west by a north-south ridge, the Tihemboka arch, to the north by the Gargaf arch, to the east by the Tibesti-Sirte arch, and to the south by the Precambrian basement of the Sahara (Fig. 2.1).

The Murzuq basin contains some 5 km of sedimentary fill mostly of Palaeozoic age (Thomas, 1995). Along the southern margin of the basin the Precambrian basement is unconformably overlain by continental red beds termed the Mourizidie Formation (Jacque, 1962). This is attributed to the Infra-Cambrian and correlated with the Purpre d"Ahnet of Algeria. The Mourizidie Formation is followed by the typical pan-Saharan Palaeozoic sequence (Bellini and Massa, 1980). This crops out in a series of intermittent subconcentric escarpments around the basin margins. All the main formations are present. On the north flank of the basin, at Wadi el Shatti on the southern limb of the Gargaf arch, there are sedimentary iron ore deposits. These are magnetite Oolitic beds within the Upper Devonian Aouinet Ouenine Formation (Turk *et al.,* 1980).

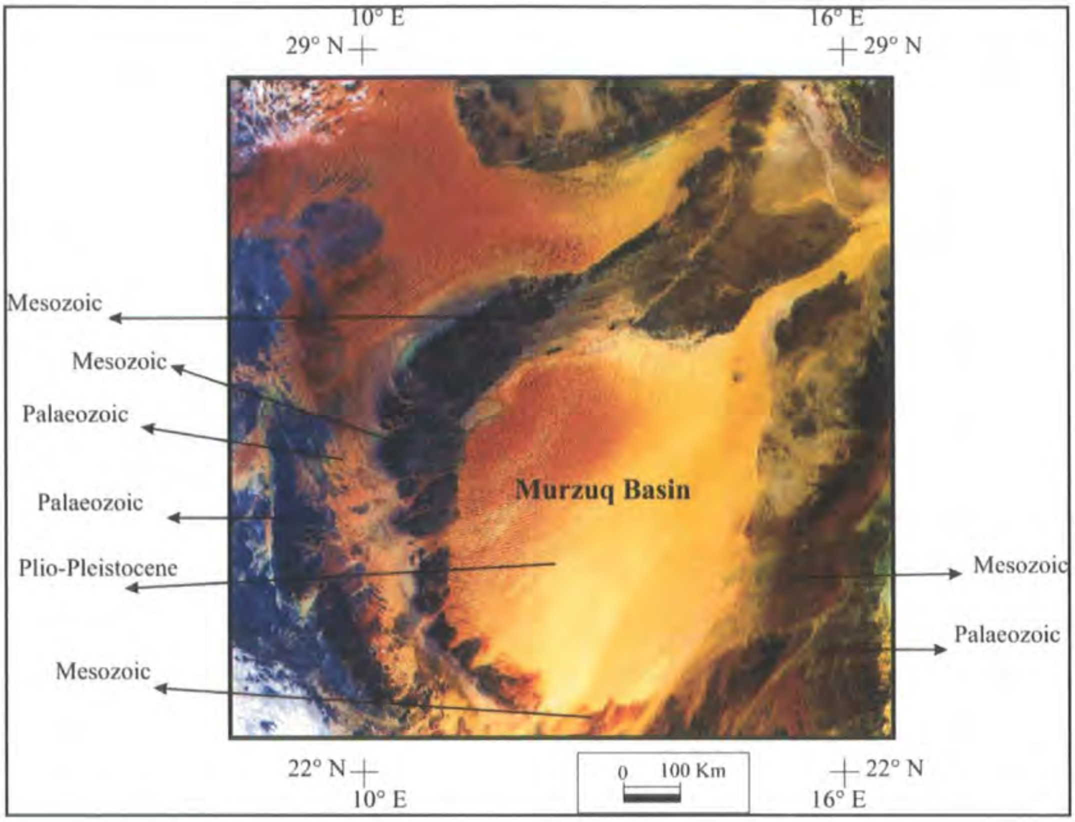


Fig. 2.5. Location map of Murzuq Basin. (From the ESA web page, www.esa.int/esaEO/index.html).

The most dramatic escarpment of the Murzuq basin rim is provided by the Continental Mesozoic sandstones, known locally as the Messak Sandstone. This takes its name from the Messak escarpment on the northern flank of the basin (Klitzsch and Baird, 1969). This formation may attain a maximum thickness of over 1,500 m in the basin centre. The age of these barren continental sediments is typically hard to establish. Spores of Triassic-Jurassic age on the southwestern flank of the basin (Pierobon, 1991). These palaeontological data imply a prolonged, if intermittent, history of continental deposition throughout the Mesozoic Era (Wealden) age (Klitzsch and Baird, 1969). The basin center contains an indeterminate fill of Holocene eolian dunes and Pleistocene alluvium.

Palaeocurrent studies show that, like the Algerian basins to the west, the Murzuq basin was once a northerly plunging embayment that opened to the Tethyan Ocean. On the Gargaf arch, the structure that defines the present northern limit of the basin, the Melez Chogranne Shale Formation (Ordovician) is locally cut out by the Mamuniyat Sandstone Formation (Ordovician). Similarly the Tannezuft Shale Formation (Silurian) cross-cuts both of these to unconformably overly the Hassounah Formation (Cambro-

Ordovician, Collomb, 1962). These unconformities show that the Gargaf arch was an intermittently positive feature in the Early Palaeozoic.

2.2.5 Al Kufrah basin (Fig. 2.6)

The Al Kufrah basin (Fig. 2.6) occupies a large part of south east Libya, though it extends northeastwards into Egypt, and southeast into Sudan and southwest into Chad. The Al Kufrah basin is separated from the Murzuq basin to the west by the Tibesti - Sirte arch and from the Sirte basin to the north by the Calanscio arch. Its eastern limb forms the western edge of the Arab-Nubian shield to the east. The Precambrian Ennedi craton defines its southern rim (Fig. 2.1).

The Al Kufrah basin (Fig. 2.6) is one of the least accessible and least known of all the Saharan basins. Nonetheless there are published accounts of its Palaeozoic sediments, notably by Turner (1980, 1991), and of its continental Mesozoic sediments, notably by Van Houten (1980) and Klitzsch and Squyres (1990). The Palaeozoic sediments are broadly comparable in facies and in their gross stratigraphic sequence to that further west. Precambrian basement is unconformably overlain by braided alluvial cross – bedded sands comparable to those of the Hasawnah Formation. They are commonly interbedded with Tigillites sands of typical Hawaz type. It is tempting to attempt to recognize the Mamuniyat and Melez Shuqran Formations of western Libya in the Kufrah basin but, though the facies are comparable, lithostratigraphic continuity appears not exist (Hallett 2002).

The centre of the Al Kufrah basin is infilled with continental Mesozoic clastics (Klitzsch and Squyres, 1990). These are broadly comparable in facies to those of the Murzuq basin. Bellini *et al.,* (1991) recognized three units of continental Mesozoic sediments. The lowest is composed of typical fluvial sands and shale's. They apply the Algerian term "Continental Post –Tassilian" to these sediments, and suggest a "Permian to Jurassic "age, citing the existence of a rich Permian pollen assemblage in borehole samples. This is overlain by a curious silicified limestone, that they term the Chieun Formation. This locally contains abundant Hyrobia, indicating a lacustrine environment, and a post –Triassic age. The topmost continental Mesozoic sediments they describe as "Nubian sandstone" and suggest an Early Cretaceous age. Similarly Klitzsch and Squyres (1990) in their synthesis of the Nubian sandstones of northeastern Africa offer no age for the continental Mesozoic sediments of the Kufrah basin beyond "undifferentiated Jurassic to Cretaceous".

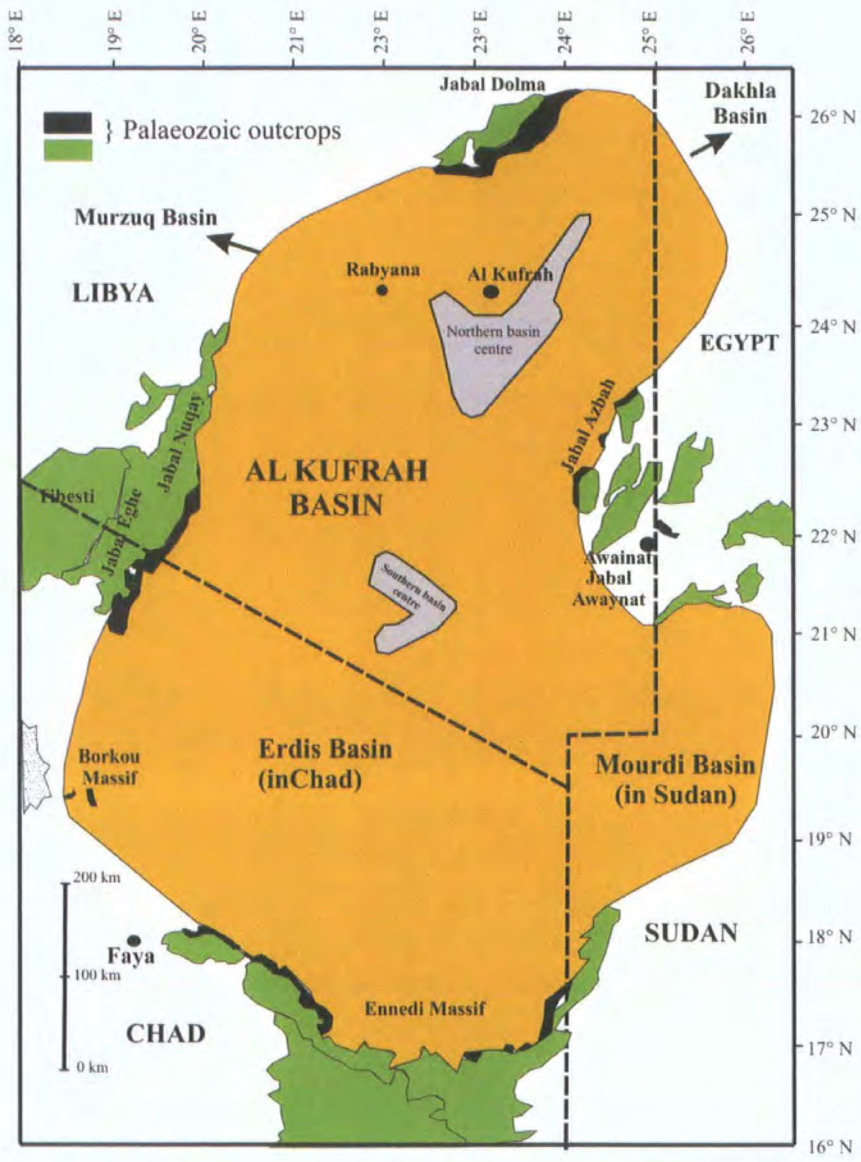


Fig. 2.6. Location map of Al Kufrah Basin. (Adapted from Hallett, 2002).

Thickness of the various formations of the Al Kufrah basin has been published from surface sections measured around the basin margin. With negligible oil exploration in the basin little is known about the thicknesses of the formations in the basin centre, and still less is publicly available. Using a combination of surface and subsurfaces data Bellini et al. (1991) suggest that there is a total thickness of some 3,500 m of sediment in the Kufrah basin. The structure and tectonic history of the Kufrah basin is also little known.

The Kufrah basin shares many features with the basins to the west. These include a similarity of facies and stratigraphy, together with predominantly northerly paleocurrent in fluvial sediments. These features all suggest that the Al Kufrah basin formed part of the undifferentiated Sahara platform, from the Cambrian until at least the Cretaceous Period. Thus the present synclinal shape of the Kufrah basin most probably postdates the deposition of the sediments with which it is filled, and was coeval with the break up of Africa and the collapse of the Tibesti-Sirte arch in the mid-Cretaceous. Trans-Saharan facies belts tend to show southwest-northeast trends. The more continental aspect of the facies in the Al Kufrah basin compared with those of the western basins suggests that it lay further from the shores of the Tethyan Ocean.

2.3. The Jifarah region NW Libya

2.3.1. Tectonic framework of Jifarah region

The previous sections have outlined the broad basinal history of Libya and the importance of Triassic Stratigraphy in many of the basin (e.g. Ghadames). The Jifarah region of NW Libya contains several well exposed sections of Late Triassic Stratigraphy that have largely been disregarded except for a few tectonic studies of the area (Anketell and Ghellali, 1991; Anketell, 1996). This section of the chapter will concentrate on the long-term tectonic framework of the Jifarah region and particularly how the Triassic Stratigraphy is affected, this is of particular importance for the Abu Shaybah Formation that forms the focus of chapters 3, 4 and 5.

The region is structurally complex affected by both the Caledonian and Hercynian orogeny (Mikbel, 1977; Goudarzi, 1980). Previous studies of the Jifarah region have concentrated on the tectonic framework between the Sirt and Ghadames basins. Both basins are of importance for hydrocarbon exploration and this intervening region assists understanding of the tectonic evolution of the basins. However, long-term tectonic activity has affected the preservation of the Triassic successions and then can only be determinate from detailed architecture studies that afford on understanding of the relationship to the associated structures.

One of the eastern extensions of the South Atlas fault system is the Sabratah – Cyrenaica fault zone (SCFZ) that extends in an ESE direction off the Libyan coast to enter onshore at the Gulf of Sirt to extend and form the boundary between Cyrenaica platform and the eastern extension of Sirt basin (Fig. 2.7). The WNW – ESE orientation of the South Atlas fault extensions in the Jifarah area is largely controlled by the Hercynian orientation of the Jifarah arch. Off the NW Libyan coast the fault zone consists

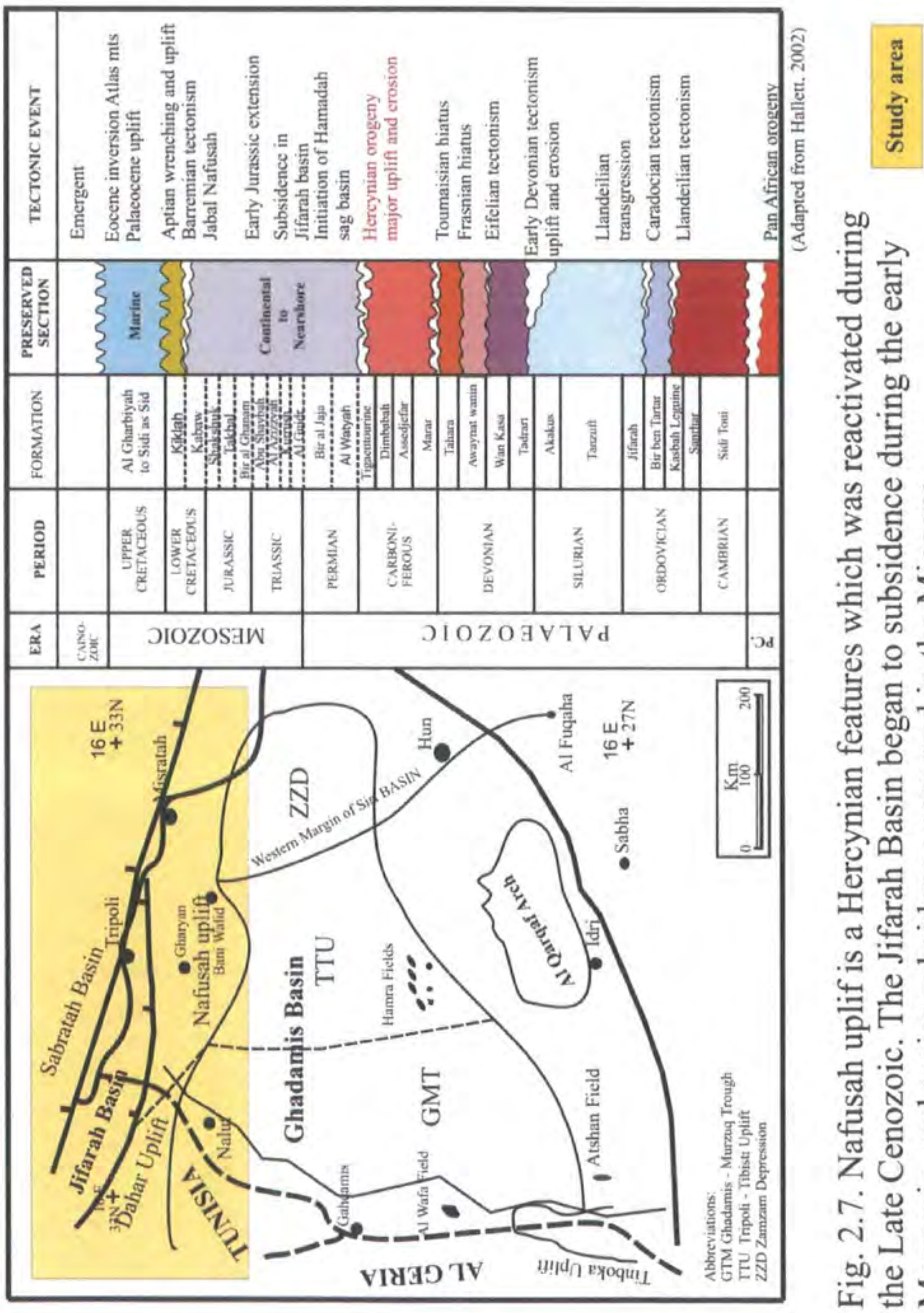


Fig. 2.7. Nafusah uplif is a Hercynian features which was reactivated during the Late Cenozoic. The Jifarah Basin began to subsidence during the early Mesozoic, and major subsidence occurred in the Miocene.

of a series of down faulted terraces on the unstable continental margin to form the southern boundary of Sabratah Basin in the Pelagian shelf. The basin movements between SCFZ and other eastbound branches of the South Atlas fault system to the north. On land to the south, the Jifarah basin's Mesozoic infill was affected by syn- depositional faulting during the Triassic and the Cretaceous, which reflect the dextral strike-slip tectonic style on the southern side of Tethys (Anketell 1996).

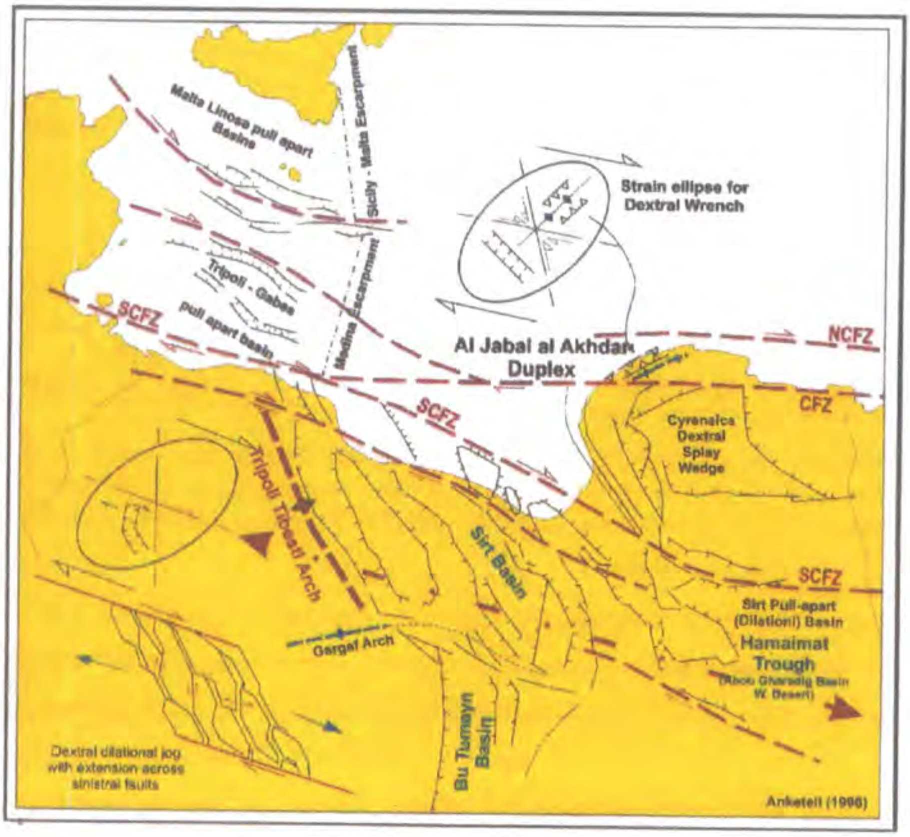


Fig. 2.8. Showing the relationship between Sabratah – Cyrenaica Fault Zone (SCFZ), the Jifarah fault, Sirt basin and Pelagian shelf (From Anketell, 1996).

In The Jifarah region that forms the southern margin of Sabratah basin, the East – West trending arrays of Riedel shears of the Al Aziziyah fault system are interpreted to be a result of dextral strike – slip wrenching along the South Atlas fault (Anketell and Ghellali, 1991; Anketell, 1996). This fault system is terminated eastward in Wadi Ghan area with a NNW-SSE trending dip-slip fault zone that was interpreted as dextral trailing extension imbricate fan splays (Fig. 2.8). Since Riedel shears are basically slip surfaces forming in sediments above a basement strike – slip dislocation during the early stage of deformation, Anketell (1996) used the Aziziyah – Wadi Ghan fault system as a model to interpret the origin of Sirt basin. By considering that the Sabratah- Cyrenaica wrench fault zone was developed in the overburden as a result of basement strike-slip deformation, the Sirt basin fault system represents its dextral imbricate trailing fan splay (Fig. 2.4).

Alternatively, Anketell (1996) proposes that since the SE end of the Wadi Ghan fault zone is buried under Tertiary lavas, it appears that the zone jogs eastward along another dextral transform fault zone to contact with the Hun Graben fault zone. This, one may interpret the Wadi Ghan fault zone as a pull-apart fault basin like the Hun Graben rather than a trailing extension fan. In the main Hercynian phase, the Late Westphalian-Early Permian Hercynian movements initiated the uplifting of the El Biod, Dahar and Nafusah Highs, resulted in intensive erosion of the Palaeozoic rocks, in some cases as deep as the Cambrian (Echikh,1998).

The effects of the earliest movements of the main Hercynian phase is well illustrated in the Illizi basin, e.g. over the Edjeleh structure, where Westphalian carbonates are observed to unconformably overlie subcropping Namurian units (Sonatrach-Beicip, 1975; Attar, 1987). In Mesozoic tectonic events, (Triassic – Jurassic) an extensional event affected the area in the Triassic- Liassic, related to the rifting of Tethys and the opening of the Atlantic (Guiraud 1998). This led to the development of a series of enechelon normal faults and tilted blocks, with associated volcanism, in the northwestern part of the Ghadamis Basin and southern Tunisia. These fault sets can be demonstrated to control thickness and facies changes within Triassic sediments. Peak activity probably occurred during the Lias (Hettangian). Bir Berkine. Si Fatima and Wadi el Tel tilted fault block traps were formed at this time Swire and Gashgesh, (2000). The Jifarah Trough was created by tectonic movements associated with the Hercynian Event, although later tectonic events have affected its genesis. And the Trough is outlined by the tectonic features of the Nefusah uplift to the east, the Telmzane Arch to the west, a southern Al-Aziziyah Fault Zone and a northern Coastal Fault Zone. The Jifarah Trough's structural development is integrally linked to that of the Ghadamis Basin to the south and the Sabratah Basin to the north.

Extension and subsidence continued into the Triassic and Early Jurassic (Morgan et al., 1998). North-south-trending normal faults and east-west trending transfer faults developed at this time, particularly along the North-South Axis in eastern Tunisia (Morgan et al., 1998). Clastic and carbonate sediments were deposited in the Triassic as well as evaporites due to sedimentary loading resulted in vertical migration and formation of diapirs and subsurface "salt walls" (Burollet, 1981). During the Early Jurassic, turbidities, as well as shelf and pelagic carbonates, were deposited.

During the Hercynian orogeny the northern Ghadamis Basin into a prominent east-west arch which was extensively eroded during the Permian, exposing the Precambrian core of Pharusian rocks in the area of wells A1-34 and D1-24. During the Permian the Nafusah Uplift formed a barrier between the marine sequences of the Tethys Ocean to north and the Continental Post Tassilien rocks to the south. By Triassic times the arch had been greatly reduced in elevation and Triassic, Jurassic and early Cretaceous sediments were deposited over the arch and into the Mesozoic Hamadah Basin to the south (Fig. 2.9). The arch was covered by the Cenomanian transgression and deposition of marine rocks continued until Eocene times. The Nafusah Arch was reactivated during the mid-Tertiary in response to the closing of Tethys. It was subjected to uplift accompanied by wrench faulting on the Jifarah Fault. Dextral strike-slip faulting can be seen at several locations along the line of the fault. Eocene tectonics also led to the production of basaltic sills and flows near Gharyan, and the volcanic activity has continued until recent times. A major escarpment formed along the line of the Jifarah Fault and Jurassic and Triassic rocks outcrop both on the escarpment and on the Jifarah Plain (Hallet, 2002).

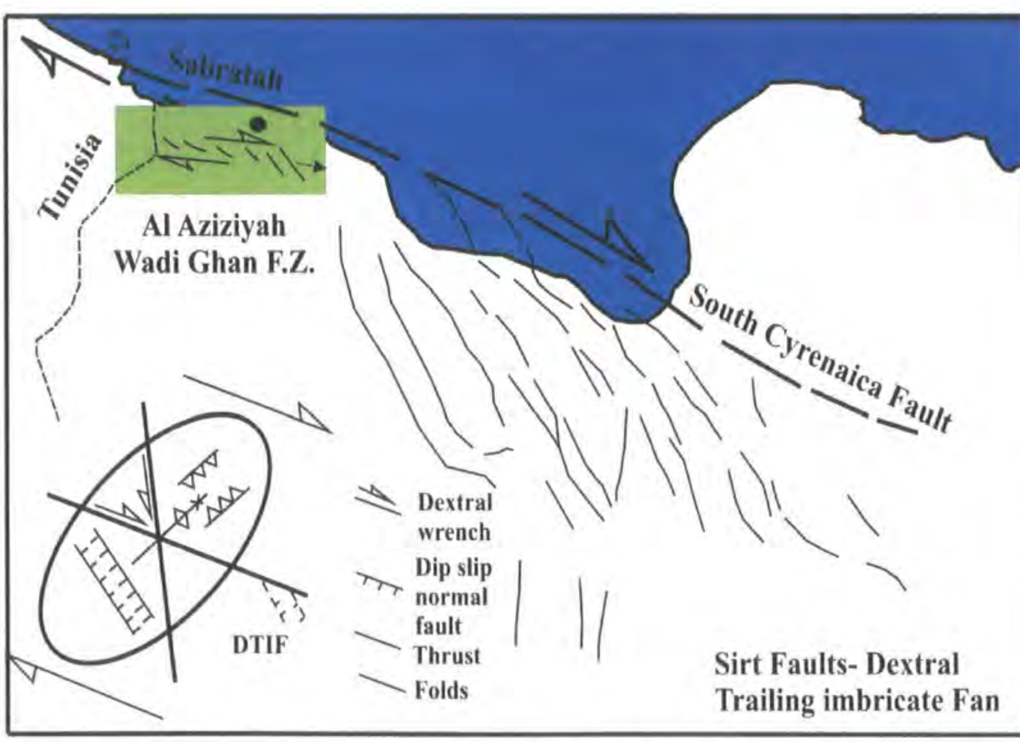


Fig. 2.9.Dextral strike slip movement along Al Aziziyah fault zone the Wadi Ghan fault system is taken as a model to interpret the relationship between the Sabratah

South Cyrenaica fault zone (SCFZ) and the Sirt rift complex (From Anketell and Ghellali, 1991; Anketell, 1996).

2.4. Early Mesozoic Stratigraphy of The Jifarah Region

2.4.1 General

Triassic rocks in Libya occur in four main domains; the predominantly marine rocks of the Nafusah escarpment and offshore, the eastern extension of the Algerian Triassic Basin into the Ghadamis area, the grabens in the subsurface of eastern Libya, and the continental rocks of the interior. Rubino et al.(2000), divided the Triassic sequence of the Jabal Nafusah escarpment into four sequences separated by unconformities. Sequence 1 sequence to the Al Guidr / Kurrush Formations, Sequence 2 to the Al Aziziyah, Sequence 3 to the Abu Shaybah Formation and sequence 4 to the Abu Ghaylan carbonates, (Table .2.1).

In the offshore the Triassic is mostly too deep to have been penetrated by wells but seismic data show thick Triassic sediments to the north of the Sabratah – Cyrenaica Fault. The Triassic rocks are well developed in the Sabratah Basin and evaporites are present in the late Triassic in the western part of the basin.

SYSTEM PERIOD	SERIES EPOCH	STAGE	TIME Ma	JIFFARAH TROUGH NW Libya (NOC)Sola et al., 1986	JIFFARAH TROUGH S. Tunisia Dridi et al., 2003	J. NEFUSA NW LIBYA Fatmi et al., 1978
JURASSIC	EARLY	Toarcian			B' hir Evaporites Series	Abu Ghaylan Fm.
		Pliensbachian				
		Sinemurian	196.5			
		Hettangian	199.6	Bir El Ghnem Formation	Messaoudi Fm.	Abu Shaybah Fm.
TRIASSIC	LATE	Rhetian	203.6			
		Norian	216.5	Abu Shaybah Fm	M'hira Fm.	
		Carnian	228.0	Al Aziziyah Fm	Al Aziziyah Fm	Al Aziziyah Fm
	MIDDLE	Ladinian	237.0	Ras Hamia Fm. Myphora Ls	Kirchaou Fm. Myphora Ls	
		Anisian	249.7	Ouled Chebbi Fm	Ouled Chebbi Fm Bir El Jaja Fm	Kurrush Fm.
	EARLY	Scythian			Bir Mastoura Fm	

Table 2.1. Stratigraphic chart of Triassic rock of the studied area.

2.4.2. Kurrush (Ra's Hamia) Formation

Description. The formation was introduced by Desio in 1960 for a series of reddish, fine grained, micaceous sandstones, silty shale's and dolomitic limestones exposed in several small domal inliers north of Gharyan. Recent remapping by IRC confirmed that the formation outcrops in a small area north of Gharyan. About 40 m is exposed but the base is not seen. The upper contact with the Al Aziziyah Formation is gradational. The formation contains a rich fauna of pelecypods, echinoderm fragments, gastropods, foraminifera, fish teeth, and bone fragments of the sauropod Nothosaurus, which allow it to be dated as Anisian / Ladinian.

The thickness of this formation is present in the subsurface of the Jifarah Basin is 581 m, a hypostratotype was designated by Hammuda et al. (2000). The basal contact with the Al Guidr Formation appears to be gradational. The maximum known thickness of 993 m.

Interpretation. The palaeoenvironment for the deposition of this unit was quite varied. In the Jifarah basin an abundance of smooth-walled trilete spores combined with the reddish brown nature of the predominantly argillaceous sediment suggests emergence and deposition in a swamp type setting. The terrestrial deposition for this interval is supported by the presence of occasional traces of coal and lignite at various horizons. Sedimentological analysis indicates a flood plain palaeoenvironment cross cut by braided and meandering fluvial systems (Dridi et al., 2003; Echikh, 1998).

2.4.3. AL Aziziyah Formation

Description. The formation is represented by massive grey dolomite and dolomitic limestone and occasional subsidiary interbedded shale. Thin anhydrite and Oolitic horizons have also been described (Hammuda et al., 2000). The age of the Aziziyah Formation is Late Triassic, Carnian, Julian to undifferentiated Late Triassic, Carnian to Rhaetian. The age is based on the highest occurrences of the miospore taxa *Infernopollenites spp., Crassicappites spp., Parillinites spp.*, and abundant Enzonalasporites vigens. In the Jifarah basin recovery of sometimes abundant miospores suggests proximity to a terrestrial flora.

Interpretation The predominance of alete bisaccate pollen (wind blown elements) at certain horizons indicates a dry hinterland and possible derivation from colder higher ground. The recovery of occasional acritachs and scolecodonts combined with the predominantly carbonate lithology suggests that deposition was in a marine inner neritic palaeoenvironment. An association of an increase in thin - walled *Leiosphaeridia spp*. With reddish brown clay and siltstone at two intervals indicates shallowing of the sea and possible emergence at these horizons (Swire and Gashgesh, 2000). A macrofauna including brachiopods, bivalves and cephalopods has also been recovered from the Al Aziziyah Formation. These confirm the Carnian age (Hammuda et al., 2000). Details of the outcrop of the Azizia Formation in the Gharyan area of NW Libya are dealt with by Assereto and Benelli (1971). Here they define six major facies which are algal biolithites, burrowed and laminated micrites and dolomitic micrites, argillaceous biomicrites and dolomitized biomicrites, pelmicrites, oosparites and medium to coarse crystalline dolomites. The depositional model for the Al Aziziyah Formation is low energy, inner platform, and shallow marine subtidal to intertidal palaeoenvironment.

The uppermost Azizia Formation exposed in the Gharyan area shows complex channeling and an erosive contact with the overlying Abu Shaybah Formation (Hammuda et al., 2000).

2.4.4. Abu Shaybah Formation

Description. The Abu Shaybah Formation in the Jifarah basin is represented by interbedded siltstones, sandstones, shale and clay with minor limestone. Outcropping lithologies in the Gharyan area are represented by red, yellow and green, fine to coarse, commonly cross-bedded sandstones, conglomerates, siltstones and shales. Dolomites and calcareous beds are reported from the top of the unit (Fatmi et al., 1978). A detailed petrography of the Abu Shaybah Formation is provided by Assereto and Benelli (1971). They divide the formation into lower and an upper member in the Gharyan area. These though cannot be recognized in the subsurface.

The age of this formation is undifferentiated Late Triassic, Carnian to Rhaetian, from palynological recovery in the Jifarah basin. Regional geological evidence suggests that the age of the interval is Norian. The Late Triassic age is based on the highest occurrences of the miospores *Patinasporites densus and Enzonalasporites* *vigens*. From outcropping areas in Gharyan the Abu Shaybah Formation show an apparent total red bed development, but pure white sandstones are obscured by run-off of red clay from the intervening claystone.

Interpretation. Fluvial braided, meandering rivers and floodplain are the main deposited environments of Abu Shaybah Formation. (Rubino et al., 2000).

2.4.5. Abu Ghaylan Formation (Jurassic rocks)

Description. The formation is composed mainly of light colored white to buff and light grey to light brown limestone and dolomitic limestone. Laminated dolomitic micrites, pelletal micrites, stromatolite horizons Oolitic and algal limestone, dolomitic breccia, gypsiferous marl and clay. The lower part is composed of fine to coarse dolomitic breccia interbedded with dolomicrite and gypsiferous clay showing thin to medium bedding. In the lower parts of the formation undulating to contorted interbeds of carbonate and marl with ripple marks, mud cracks, solution collapse breccias and tepee structures have been noted. The contorted bedding has been related to syn-sedimentary tectonic influences (Hammuda et al., 2000).

Fossils are rare in the Abu Ghaylan Formation although fish and fragments of the bivalve Astarte have been recorded from sandy lag deposits at the formation base (Hammuda et al., 2000). The middle to upper part is medium to thick bedded massive with Oolitic interbeds in the middle. They form wavy bedding which is very typical of this formation.

Interpretation. Asserto and Benelli (1971) studied the sedimentology and recognized a number of low energy supratidal to subtidal facies including laminated dolomitic micrites, pelletal micrites, stromatolite horizons and solution-collapse breccias (Hammuda et al., 2000).

2.5. Depositional Model of Triassic for the Jifarah Basin

The Abu Shaybah Formation represents a regressive low stand deposit that was deposited during a eustatic sea level fall (Swire and Ghashgesh, 2000) (Fig. 2.10). The period of extensional tectonism and rifting that was prevalent in the region from Middle to early Late Triassic also came to an end. The result was the infilling of the Jifarah Trough with an evaporites sequence. Asserto and Benelli (1971) suggest a shallow marine, inner neritic palaeoenvironment for the lower part of the Abu Shaybah Formation passing upwards to transitional and non-marine deposition. Cross bedded channel sandstones with erosive bases containing lag deposits have been recorded from outcrop in the Gharyan area (Hammuda et al., 2000).

The vertical evolution of the fluvial system in a single sequence records a strong decrease of the slope angle of the fluvial profile, braided rivers being associated with a steeper graded profile than meandering complexes (Rubino, et al, 2000). This succession fits very well with the sequence stratigraphic model developed by Jervey (1988) and Posamentier et al. (1988a, b) that predicts that braided rivers are predominantly active in the late low stand phase (transgression phase) commonly infilling the incised valley produced during that phase. Then during the transgression, the slope of the fluvial profile exceeding the minimum value for the river to flow and because of the landward shift of the coastline, bypass commonly occurs. Finally when progradation starts in a shallow marine setting, coeval fluvial aggradations leads to the development of meandering channels on a flood plain.

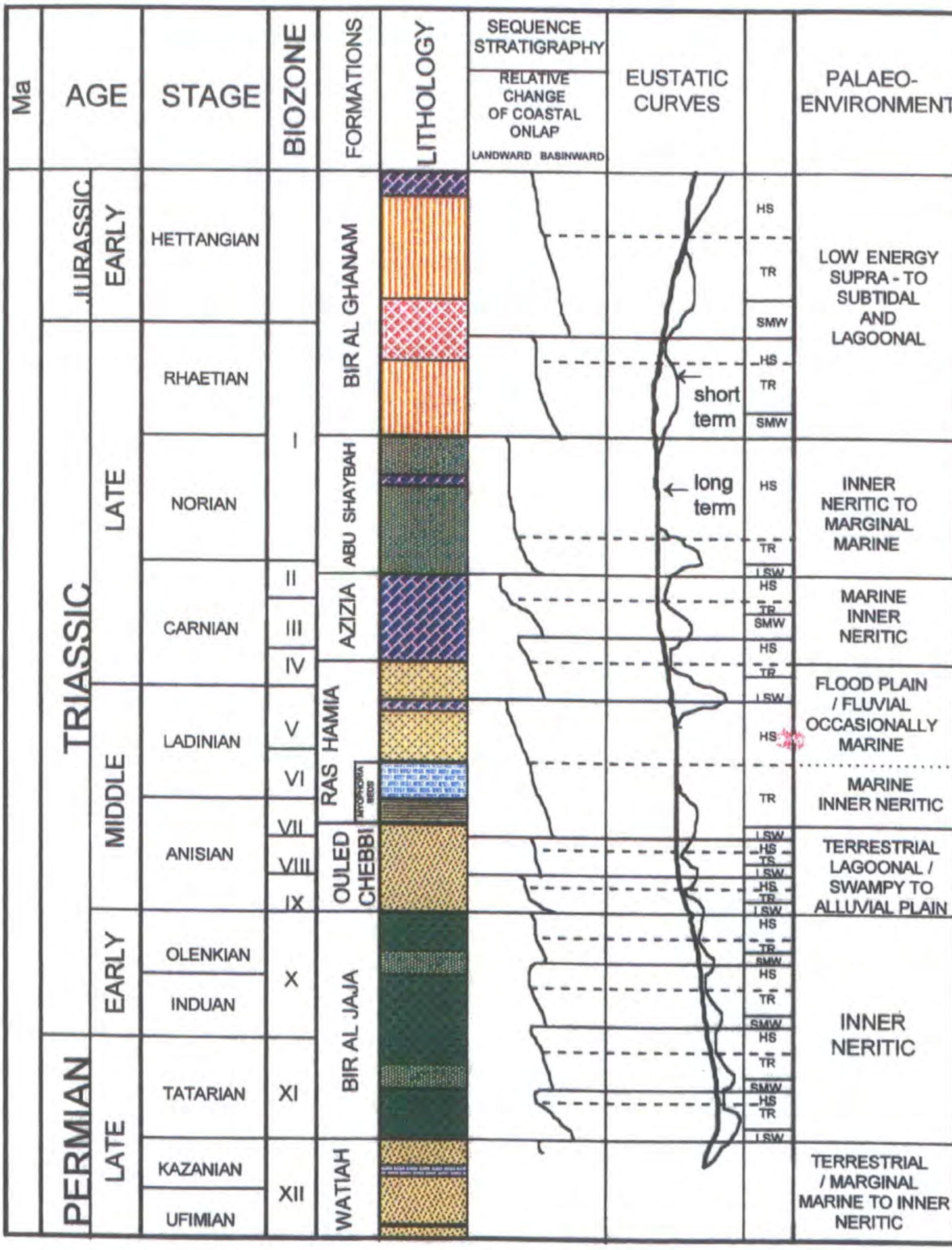


Fig. 2.10. Sequence stratigraphy of the Triassic section for the Jifarah region, NW Libya From (Swire and Gashgesh, 2000).

2.6 Summary of key events

Triassic in North Africa was associated with the early phases of rifting prior to the break-up of the supercontinent of Pangaea some 220-175 M.a. Rifting and associated magmatic activity of Pangaea followed the Hercynian event and continued through the Triassic and well into the Jurassic time. The Neo-Tethys was opened during the Middle Jurassic along a triple-point junction that separated Europe from Africa. These events coincided with the initial Atlantic rifting which continued during the Mesozoic.

The Pangaea land-mass was also profoundly influenced by the prevailing palaeoclimatic conditions. The area of the present day North America, Western Europe and North Africa was positioned within a broad equatorial belt characterised by year round arid climate that promoted sediment oxidation and led to the development of red beds. The distribution of the Triassic rift basins of these areas and their infill of red bed facies was located approximately along the same palaeolatitude 10° North. Many of these continental red beds were deposited as alluvial fans, fluvial braided and meandering stream systems and aeolian deposits which were often associated with sabkha, playa, lacustrine, and evaporitic facies.

These facies are widely distributed across Libyan in many of the sedimentary basins e.g. Jifarah, Ghadames, Al Kufrah and Sirt Basins.

Over 1.5 km of Triassic sediments can be found in the Jifarah Basin and thins eastwards into the Sirt basin where < 500 m occurs. The Jifarah Basin began to subside during the early Mesozoic and Jabal Nafusah Uplift by erosion during Hercynian orogeny.

This chapter has described the importance of tectonic and stratigraphic framework of Triassic rocks across the major North African sedimentary basins, but with special emphasis on Libya. The following chapter will focus on the fluvial sediments of Abu Shaybah Formation (Late Triassic) that is well exposed along the Jabal Nafusah escarpment and provides one of the first detailed facies analyses of Late Triassic fluvial red beds that is contemporaneous with shallow marine sedimentation of North Africa and Europe. Such a difference in sedimentary facies provides an ideal setting for understanding climatic controls on fluvial systems and the sequence stratigraphy of non-marine sequences.

Chapter 3

Architecture and Evolution of Triassic Fluvial System, Abu Shaybah Formation, NW Libya

Chapter 3

Architecture and Evolution of Triassic Fluvial System, Abu Shaybah Formation, NW Libya

3.1 Outline of Triassic research

The Triassic in a worldwide stage has long been recognised as an important geological period due to the predominance of fluvial red beds and their potential to act as petroleum reservoirs and aquifers. Exploration in Libya has somewhat neglected the Triassic unlike neighboring North Africa countries (e.g. Algeria TAG-I). However in recent years there has been an awareness of the importance of the Triassic in the Ghadames basin, and potential findings in the Sirt and Al Kufrah Basins (See chapter 1 and 2). Much of the Triassic succession in Libya is only found in the subsurface and most previous knowledge is based on seismic and well logged sections. The Abu Shaybah Formation (ASF) of the Jabal Nafusah area produce one of the few well exposed sections of Late Triassic fluvial successions (Fig. 3.1). This chapter will focus upon the architectural elemental analyses recognized and well defined geometries and bounding surfaces using the methodology of Miall (1996). Fluvial strata provides an objective method for determining the three dimensional geometry of ancient river systems (Allen, 1983; Hirst, 1991; Miall, 1993, 1994). There are, however, relatively few examples from detailed field observations of the three-dimensional geometry of an entire braided to meandering fluvial system along the escarpment of Jabal Nafusah. The ASF offers ideal conditions for a detailed architectural study of a braided and meandering fluvial system. The excellent outcrops, the large size and well- established geology allows for detailed analysis of the architecture in the context of structural control and along-strike sediment supply variations.

This chapter focuses on the Late Triassic sediments of the ASF which was deposited during the Hercynian orogeny in incised valley systems then filled by a thick sequence of stacked amalgamated fluvial braided channel sandstones, meandering and flood plain deposits. This incised valley was formed by both erosional and tectonic force and changes laterally into evaporites of the Bir El Ghanam Formation (Late Triassic - Early Jurassic) in the west of Gharyan (Fig. 3.1). These sediments provide a good example for understanding controlled climatically dry, wet and humid seasonal variations during deposition.

3.2 Geological setting

The northwestern part of Libya is comprised of two major physiographic units; the Jifarah basin which is bounded to the north by the Mediterranean Sea and, the escarpment of Jabal Nafusah which forms a prominent escarpment bounding the plain to the south. In broad terms the plain is formed of Pleistocene and Holocene sands and gravels underlain by Miocene and Mesozoic rocks which are exposed in the escarpment (Anketell and Ghellali, 1991), (Fig. 3.1). Over 1.5 km thickness of siliciclastic, carbonate and evaporites sequences are exposed along a 400 km long horseshoe shaped escarpment extending from the town of Al Khums, NW Libya, westward to Gabes in Tunisia. Jabal Nafusah elevation ranges from a few meters in the east to over 800 m in the centre south of Gharyan, and if descends to 200 m in Gabes. The escarpment is a retreating fault scarp, its shape and extension is controlled by the E-W trending Al- Aziziyah fault zone and the Jifarah Arch. The Abu Shaybah Formation (ASF) of Late Triassic (Carnian to Norian age), forms the foothill slopes of the Tarhuna - Gharyan scarp stretching westwards to Ar Rabitah and Al khums to the east along the Jabal Nafusah. The Lower boundary is sharp and unconformable with the Al Aziziyah Formation, and the upper is locally unconformable with the Abu Ghaylan Formation. The maximum thickness of Abu Shaybah Fm. is about 254 m located in Wadi Ghan (Fig. 3.2).

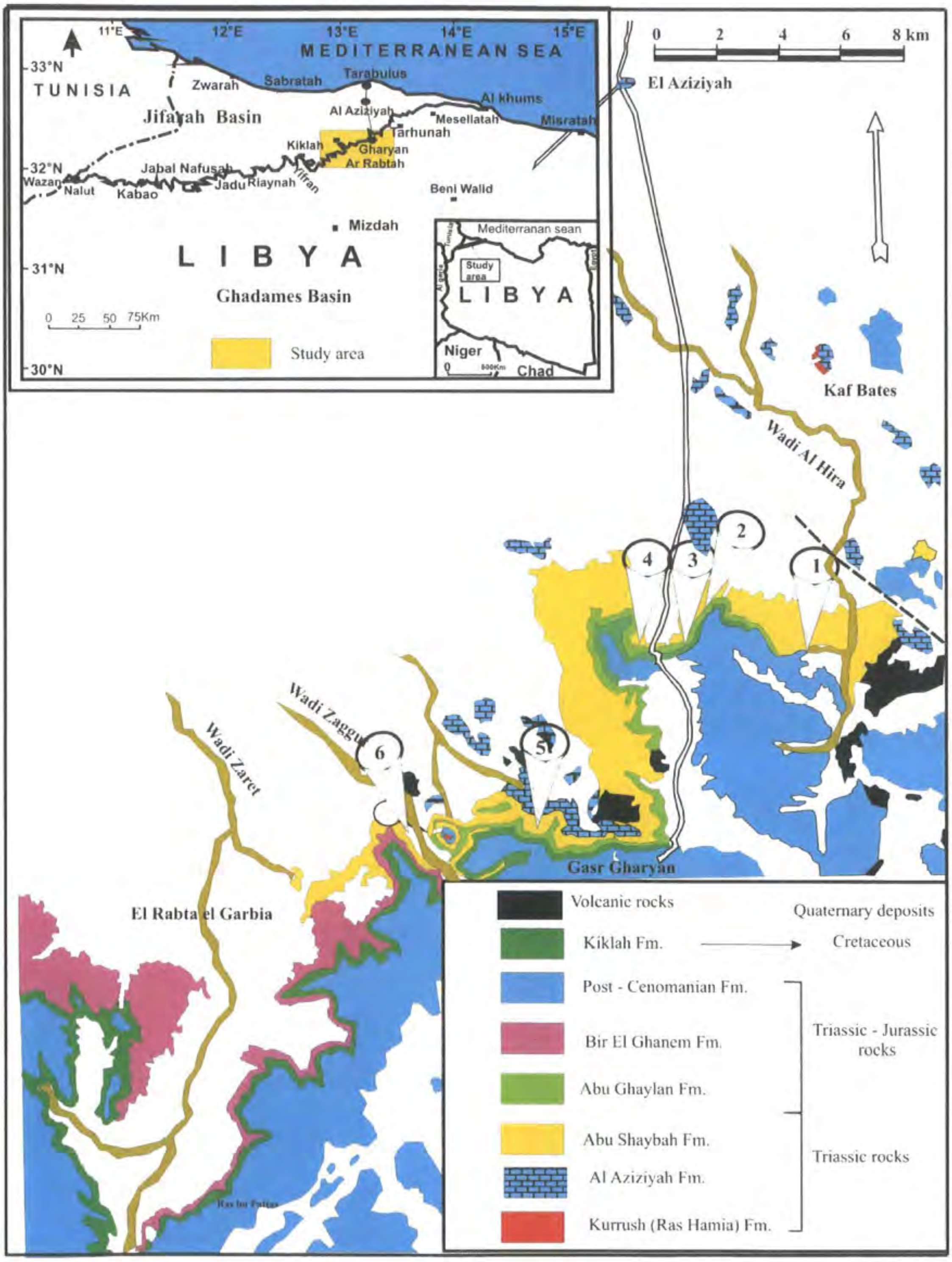


Fig. 3.1. Geological sketch of study area and location of measured sections, (1) Wadi Ghan section, (2, 3 and 4) Abu Ghaylan Road sections, (5) Gharyan dome section and (6) Abu Rashada road section (Modified from Desio *et al.*, 1963).

3.3 Sedimentology

The sediments of the ASF are dominated by gravelly coarse - grained sandstones, medium to coarse - grained sandstones, fine - grained sandstones, siltstones and mudstone and some poorly developed palaeosols. For the propose of this study the sediments have been subdivided into different lithofacies and then later grouped to form architectural elements Tables 3.1 and 3.2 (Allen, 1983; Miall, 1983, 1988a). The data for this study were primarily collected by means of outcrop facies analysis performed on six main sections located within the confines of the ASF (Fig. 3.1). The results of this work, as well as the interpretation of the data, are illustrated in Table 3.1 and 3.2.

Previous research has interpreted the Abu Shaybah Formation as a fluvial system (Rubino *et al.*, 2002). However, the fluvial system can be subdivided into braided and meandering stream deposits. Using lithofacies and facies associations this study documents the regional alluvial architecture of the ASF.

The ASF was divided into nine lithofacies (Table 3.1) which were adapted from Miall (1978). The nine lithofacies were grouped into four facies associations. Each association is described and interpreted utilizing two and three dimensional architecture found in both vertical and lateral profiles.

The studied sections show a dominantly clastic succession with lithologies ranging from claystone and shale to sandstones and conglomerate lag deposits. Numerous iron- rich palaeosoil horizons occur and assist inter regional climate interpretations for the Triassic.
The dominance of a particular lithofacies or group of lithofacies, in different parts of the Formation allows the definition of four main facies associations, as outlined below (Table 3.1).

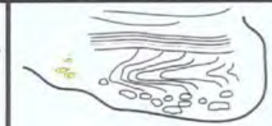
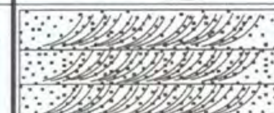
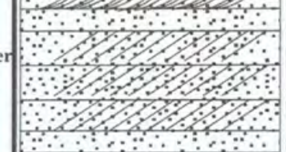
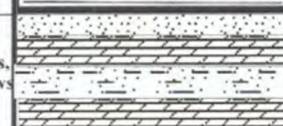
No.	Type of Facies	Description	Sedimentary structures	Interpretation	Thickness (m)
1	**Facies 1a. Conglomerate channel (Gsh)**	**Horizontally cross stratification gravels, laterally, and mud intraclast up to .5cm in diameters, scour surfaces (Se).**		Channel deposit (CH, Se)	0.1 - 2m
	Facies 1b. Planner cross strati. (Gsp)	**Coarse to very coarse grained sandstone, normal gradded planner cross stratification.**		Channel deposit (CH, Se)	0.1 -1.5m
	Facies 1c. Trough cross strati. (Gst)	**Coarse to very coarse grained sandstone, normal graded trough cross stratification.**			
2	**Facies 2a. Trough cross strati. (St)**	**Sandstone, medium to coarse grained, pebbly scattered, trough cross stratification (St).**		Crevasse Spaly deposit	0.1 -0.5m
	Facies 2b. Planner cross strati. (Sp)	**Sandstone, medium to coarse grained, pebbly scattered, planner cross stratification (Sp).**		Point bar and/ or channel deposit (SS)	0.1 -1.5m
	Facies 2c. Ripple cross strati. (Sr)	**Sandstone, fine to medium grained, horizontal cross stratification, and current ripples.**		Point bar and/ or channel deposit (SS).	0.1 -1.5m
3	**Facies 3a. Siltstone - mudstone (Sl)** **Facies 3b. Mudstone (Fsc)**	**Clay to silt in size, and sand scattered. Stratification .**		Over bank and flood plain.	0.1 -25m
4	**Facies 4a. Wavy ripples (Sr)**	**Sandstone, very fine grained inter- -calation with clay and dolomite beds, wavy ripples abundant by H. Burrows**		**Transitional zone**	20m

Table 3.1. Facies associations for the Abu Shaybah Formation, (ASF). Facies code adapted from Mail (1978, 1985, and 1996).

No	Intervals	Lithofacies code	Lithofacies description	Grain size	Thick.
5	(154.5 – 221.5m)	**Sand sheet (Ss), Over bank (OF)** Se, St	Ss with clay intervals, - fining upward cycles, - moderate to good sorting, trough and planer cross - stratification.	Clay, M-C Ss	Total thickness is **67m** (clay thick about 45m, sandstone thickness about 22m)
4	(132.5 -154.5m)	**Channel – fill complex (CH)** Se, Gsp, Gst, St, Sr and Se	Ss, small and large scale of trough cross-stratifications, climbing ripples, convolute structures and at the base of channel abundant clasts.	Coarse Sand sheet	Total thickness is **22m** (Single channel 2 -2.5 m thick.) No clay in this interval
3	(52 – 132.5m)	**Sand sheet (Ss)** **Over bank (OF)** Se, St, Sr	Ss with clay intervals, - fining upward cycles, - coarsening in small scale moderate to good sorting, trough and planar Xstrata. and convolute structure.	Clay, M-C Ss Clay- is reddish colour.	Total thickness is **78m** (Single channel 0.5 -1.5 m thick.) Clay thick about 53.5m Sandstone thickness about 22.5m Mix clay with sandstone about 2m
2	(34 – 52m)	**Channel – fill complex (CH)** Se, Gsp, Gst, St and Se	Ss, ferruginous, very poorly sorted, graded trough and planar x. strata.. Clasts up 10-15cm in diameters at the base of channel	Gravel, C-VC Ss	Total thickness is **18m** (Single channel 0.5 -2.5 m thick.) No clay in this interval.
1	(0 – 34m)	**Flood plain** Fl	Mudstone, vary coloured, calcareous.	Mud	Total thickness is **34m**

Table 3.2. Summary of facies association and main characterization of each lithofacies in the Abu Shaybah Formation (ASF), Wadi Ghan section.

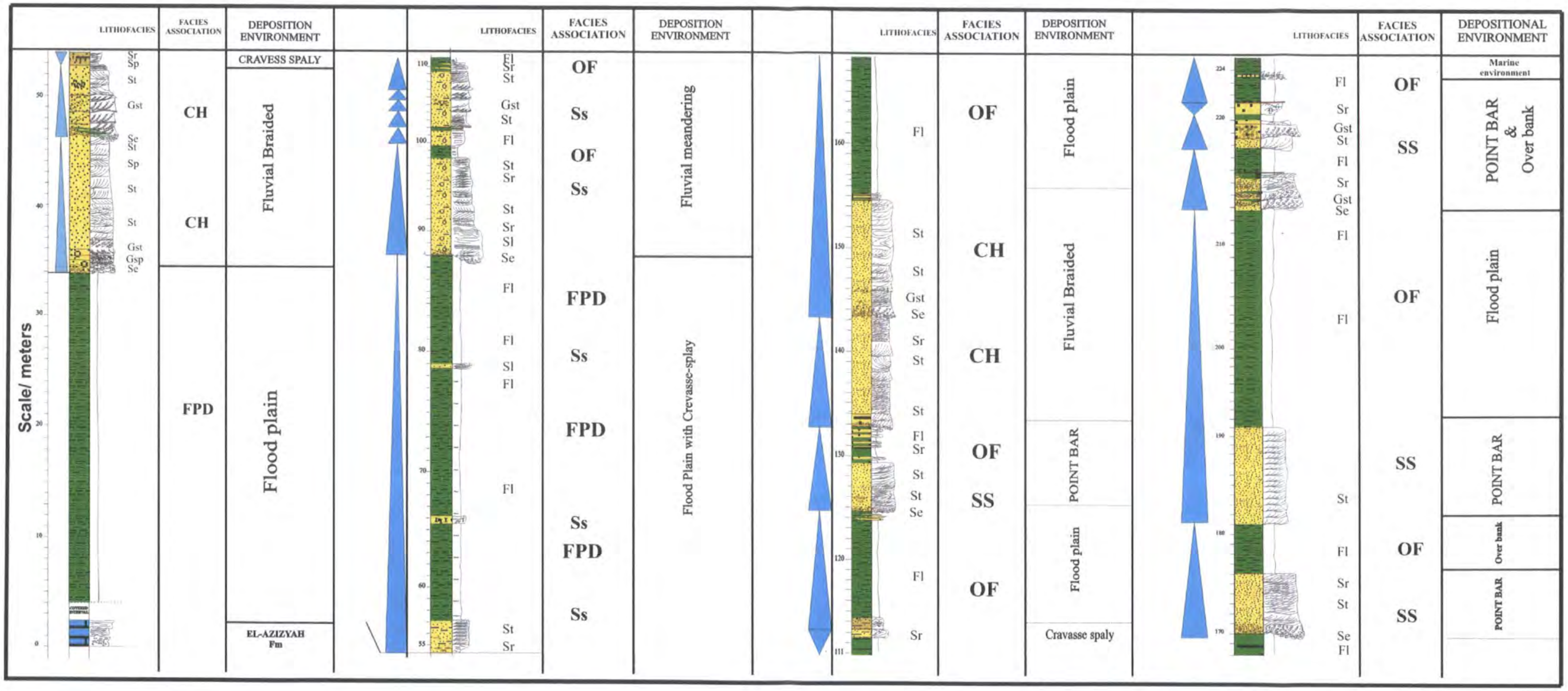


Fig.3.2. Detailed log of Abu Shaybah Formation (ASF) using architecture elements in fluvial systems.

3.3.1 Facies association

Detailed analyses of the sedimentology (i.e. grain - size, sedimentary structures, body and trace fossils) have allowed seven main facies association to be identified within the study area (Fig. 3.1). These were deposited within four main depositional environments: fluvial braided, fluvial meandering, flood plain and transitional marine.

Facies Association 1: Fluvial braided (FB) (Figs. 3.2 and 3.3)

This facies association comprises sandstone, coarse to very coarse-grained facies that were deposited within a fluvial braided system. The facies is formed of approximately 70% sandstone, 30% conglomerate.

Facies 1a. Conglomerate channel

Description. This facies is interbedded with lithofacies 1b and 1c and represents less than 30% of the total thickness (about 3 m thick and limited lateral extent to about 25m.). It is predominantly composed of coarse to very coarse-grained sandstone, cross bedded, as single sets, or more often in multi-storey associations (Figs 3.3 A - F). Multi-storey cross–bedded, sandstones of this lithofacies are often continuous for several hundred meters. However, some examples are laterally truncated or grade into other facies over distances of a few tens of meters. Sharp boundaries between individual cross bed sets and often marked by thin but persistent siltstone partings (Fig. 3.3 D). Facies 1a., typically grades laterally and vertically into smaller scale bed packages of facies 1b or 1c and rare siltstone and mudstone found as mud matrix in this facies. Trough cross-bed is compound sets generally 0.3-0.5 m thick (rarely up to 1 m) but multi-storey suites reach up to 10 m in thickness. Set widths are up to 5 m. Grain-size are generally bimodal within this facies although a pebble layer is sometimes present at the base of cross–bed sets and a slight reduction in grain-size to medium-grained sandstone normal graded cross stratification. Also containing intraclast at the base upto 10 – 20 cm in diameter, are relatively common near the base of multi-storey size, decreases upwards within up section in diameter. Soft-sediment deformation is common in the form of large-scale convoluted bedding (Fig. 3.3 C).

Interpretation. Lithofacies **1a** is essentially identical to the massive to horizontal cross stratification , trough and planar cross-sets of Miall (1977, 1996) although bedforms reach slightly greater dimensions in the lower part of the ASF in individual cross bed sets may occasionally reach 1 m in thickness and cosets may reach thicknesses of up to 10 m. The predominance of pebbly and trough cross-bedding in facies 1a suggests deposition within major channels as large migrating dunes (megaripples of Singh & Kumar, 1974). The small size of extraformational clasts and relative frequency of siltstone intraclast above scour surfaces suggests that deposition occurred at a considerable distance from the chief sediment source area or that the sediments have been largely reworked from older, mature, sandstone-dominated sedimentary sequences.

Paleocurrent for facies 1a chiefly reflect from east to west directed sediment transport in fluvial braided unit 1.

Facies 1b. Planar cross stratified sandstone (Gsp)

Description. This facies consists of coarse to very coarse-grained sandstone with scattered pebbles. Planar (Gsp) cross-bed sets are generally 0.5 – 1.5 m thick but multi-storey sets reach 10-13 m in thickness. Set widths are up to 20 m. Grain-sizes are generally relatively uniform within this facies although a pebble scattered from quartz grains, white colour, silica, carbonate and ferruginous cemented. (Figs. 3.3.E and F)

These sets occasionally occur in multi-story packages up to 3 m thick. Sp facies suites are sheet – like, ribbon- like or broadly lenticular and extend laterally for several tens of meters. They lens out laterally or are truncated by low-angle erosion surfaces into wedge-shaped terminations. This facies generally rest sharply on other association sandstone facies.

Paleocurrent foresets azimuths commonly show flow to the west direction of cross stratification measurements.

Interpretation. This facies 1b corresponds to Miall's (1977) planar cross –bedded sandstone facies (facies Sp) and Rust's (1978) low – angle cross- stratified sandstone facies (facies Sl). Medium to large-scale bedforms producing planer cross bedding in modern sandy braided rivers include both linguoid and straight-crested transverse bars (Smith, 1971, 1972). Low angle cross-beds may represent deposition under high energy conditions or in special channel location such as the base of longitudinal bars at the intersections of channels.

Facies 1c. Trough crossed-stratified sandstone (Gst)

Description. This facies consists of mostly coarse to very coarse-grained, cross bedded, sandstone as single sets, or more often in multi-storey associations. Trough cross-bed sets generally 0.3-0.5 m thick rarely up to 1.5 m, but multi-storey suites reach 2 m in thickness. Set widths are up to 200 m. Grain-sizes are generally relatively uniform within this facies although a pebble scattered at the base of cross- bed sets and a slight reduction in grain size to medium-grained sandstone is sometimes evident. Siltstone intraclast, up to 10-15 cm in diameter, are relatively common at the base of multi-storey sandstones suites. Soft sediment deformation is present in the form of large-scale trough cross stratification (Fig. 3.3A) and convoluted bedding. (Fig. 3.3 C)

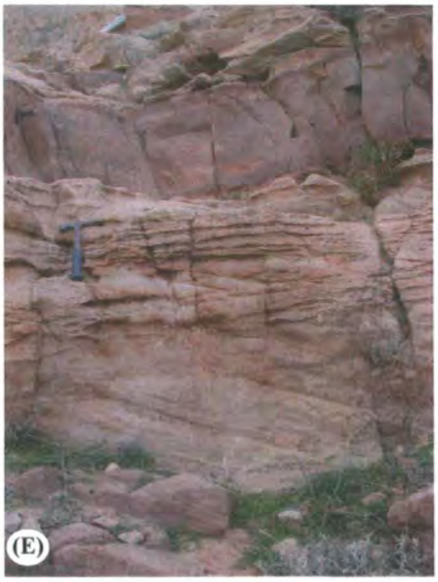
Interpretation. Facies **1c** is essentially identical to the trough cross-bedded sandstone facies (Facies St) of Miall (1977) although bedforms reach slightly greater dimensions in the ASF (individual cross bed sets may occasionally reach 1.5 m thick and cosets may reach thicknesses of up to 5 m. The predominance of festoon trough cross-bedding in (CH) Facies suggests deposition within major channels as large migrating dunes (megaripples of Singh and Kumar, 1974). The scarcity and small size of extraformational clasts and relative frequency of siltstone intraclsts above scour surfaces suggests that deposition occurred at a considerable distance from the chief sediment source area or that the sediments have been largely reworked from older, mature, sandstone dominated sedimentary sequences.

Trough cross-stratification (St).

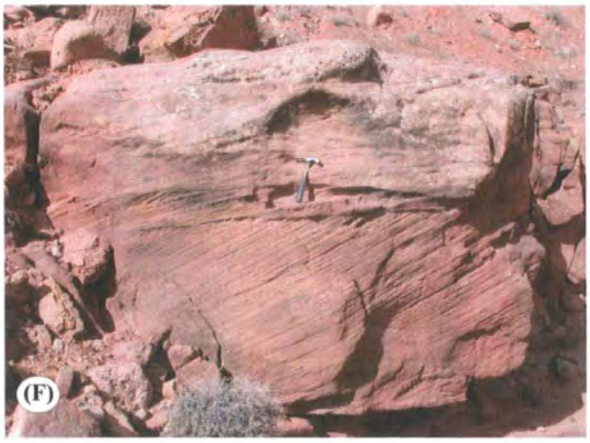
Massive pebbly sandstone, heamatitic mud matrix.

Convoluted structures in trough cross-stratification.

Gravel, trough cross-stratification (Gst).



Large and small-scale of planner cross-stratification (Gsp).



Large simple- angle planner cross-stratification (Sp).

Fig. 3.3. Facies association 1: Fluvial braided (FB), Lithofacies 1a &b, Facies association 2: Fluvial Meandering (FM), Lithofacies 2b, (photo. F)

Facies Association 2: Fluvial Meandering (FM)

Facies 2a. Trough cross stratified sandstone (St) (Fig. 3.4)

Description. This facies consists of troughs (St) develop by the migration of 3-D dunes, composed of medium to coarse-grained sandstone. Pebbles present in this lithofacies (St), clean quartzitic, moderately sorted sand and there is commonly a lag of poorly sorted sand with large intraclast at the base of the trough cross stratification consists of curved sets., ranging 1.5-2 m in thickness and extending over several ten meters laterally with paleocurrent directions related to cross bedding to the north.

Interpretation. Facies 2a (St). Troughs develop by the migration of 3-D dunes (Miall 1977, 1996). Bedforms reach slightly greater dimension in the ASF (individual cross bed sets may occasionally reach 1.5 m thickness and cosets may reach thicknesses of up to 3 m in small scale). The predominance trough cross-bedding in Facies 2a suggests deposition within major channels as large migrating dunes (megaripples of Singh & Kumar, 1974).

Facies 2b. Planar cross stratified sandstone (Sp).

Description. This facies consists of mostly coarse to very coarse grained, cross-bedded, sandstone as single sets, or more often in multi-storey associations. Trough cross-bed sets are generally 0.3-0.5 m thick, rarely 72up to 1.2 m) but multi-storey reach 2 m in thickness. Set widths are up to 200 m. Grain sizes are generally relatively uniform within this facies although a pebble layer is sometimes present at the base of cross- bed sets and a slight reduction in grain size to medium-grained sandstone) is sometimes evident. Siltstone intraclast, up to 10 -15 cm in diameter, are relatively common at the base of multi-storey sandstones suites. Soft sediment deformation is present in the form of large scale convoluted bedding.

Interpretation. Facies 2b. This facies forms by the migration of 2-D dunes (Miall, 1996). Sand is transported up the flank of the bedforms by traction and intermittent suspension (commonly forming a carpet of small-scale current ripples), and deposited at the crest, where bed-shear stress drops at the point of flow separation. Large scale cross-bedding of this facies 25° (Fig. 3.3 photo F), with sharp, angular bases, upper and lower bounding surfaces are typically flat, with little evidence of scour surfaces. Sandstone in this facies is typically sorted by the process of ripple migration up the stoss side of the dune, resulting in foresets is which the modal sand grain-size may differ by several size classes (Miall, 1996).

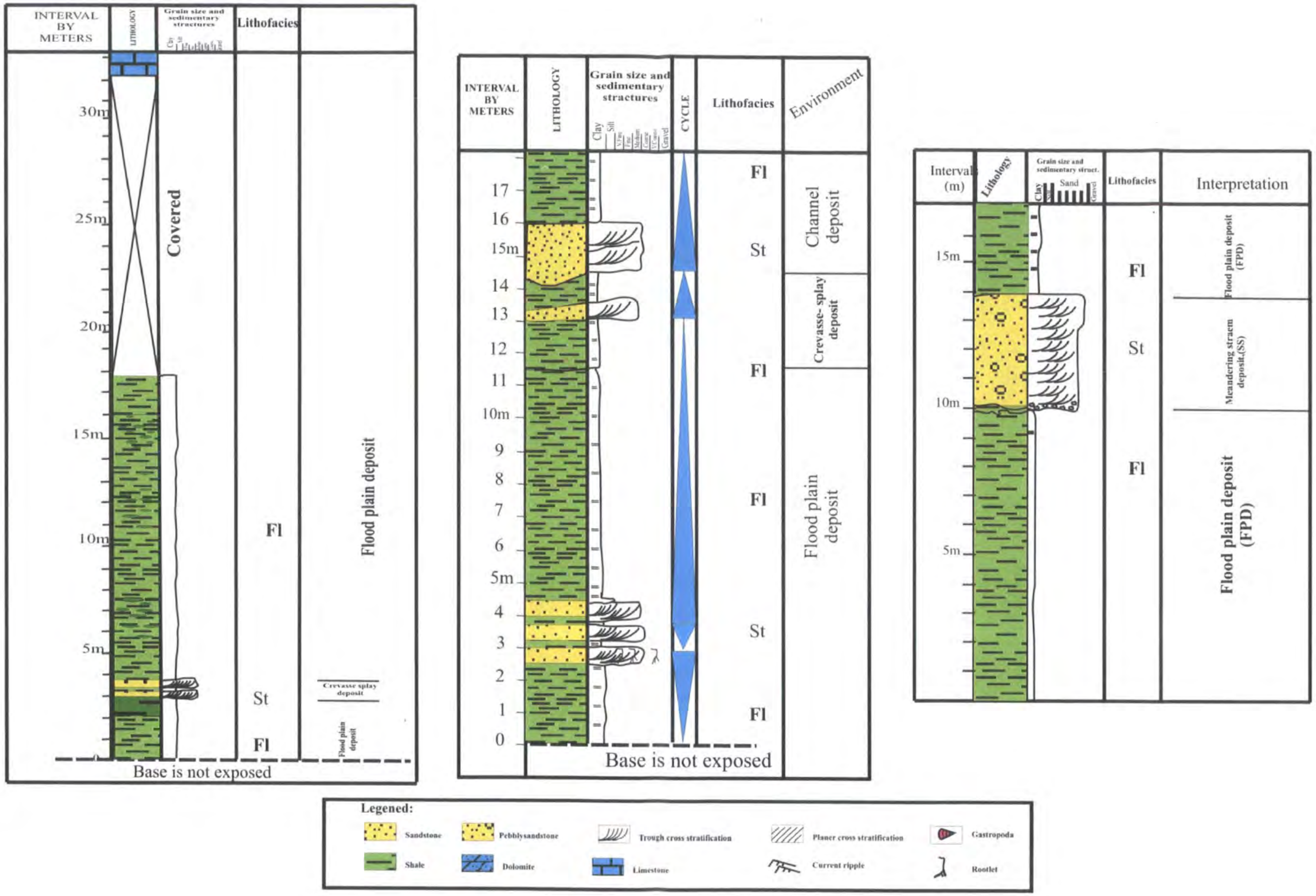


Fig. 3.4. Graphic logs of the fluvial meandering, and flood plain deposits in Abu Shaybah Formation (Localities 2, 3 and 4 in Fig. 3.1).

Facies 2c. Ripple cross-stratified (Sr).

Description. This facies consists of fine to medium-grained sandstones, trough cross-bedded and large-scale, climbing ripple (Fig. 3.5A), cross laminated sandstones. Individual cross-bed sets and cross laminated beds are typically 5 – 20 cm thick. Generally 1.5 m thickness (Figs. 3.5 A and B). A wide variety of internal structures may be generated from ripple migration, depending on flow velocity and the rate of sediment supply (Jopling and Walker 1968; Allen 1983).

Interpretation. Ripples develop at low flow speeds (< 1 m / s), and are very sensitive to changes in flow conditions (Miall 1996).

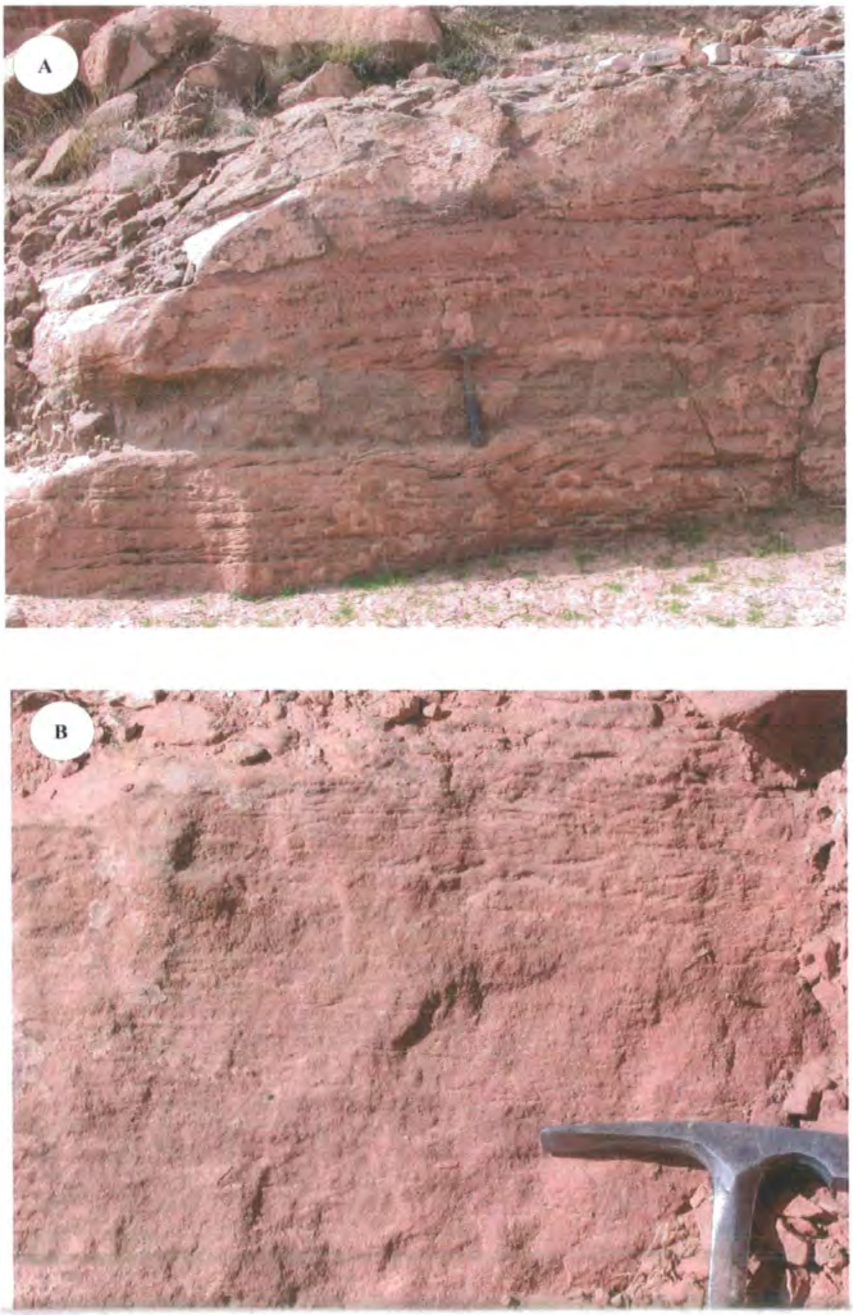


Fig. 3.5. Facies association 2: fluvial meandering (FM), Photo (A) climbing and photo (B) current ripples in Facies 2c

Facies Association 3: Flood plain deposit (FP)

Exposures in this section are composed of stacked fining-up cycles of clean sandstone, friable, trough cross-bedded sandstones grading into variegated claystones and siltstones representing meandering stream deposits. These grade from the lower sets of medium to coarse-grained sandstones highly cross planar and trough stratification (Sp & St) deposited by a braided stream (Fig. 3.6).

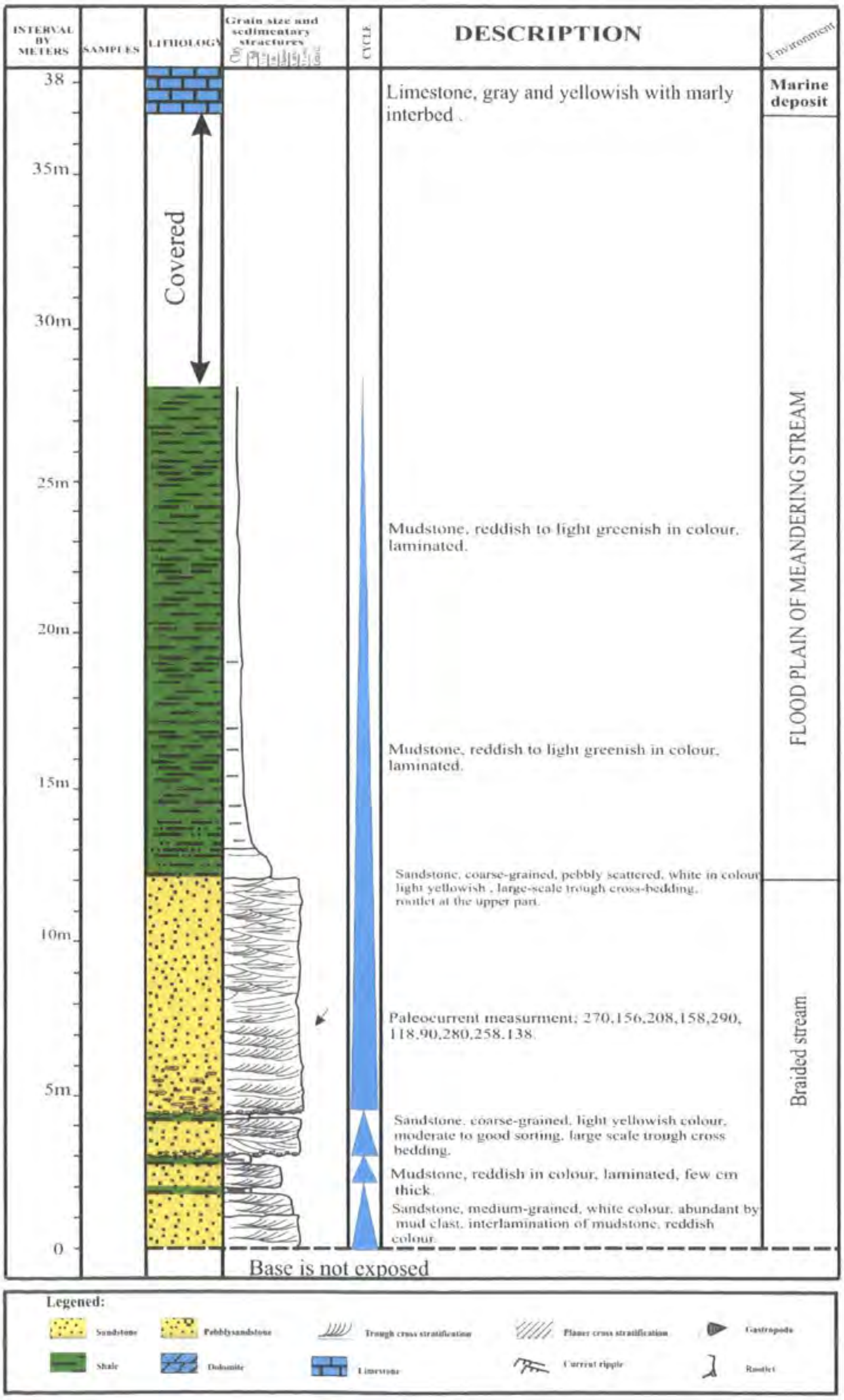


Fig. 3.6. Outcrop logged section of floodplain facies (FP).

Facies 3a. Siltstone- mudstone-fine sandstone grained facies, (Sl)

Description. This facies consist of fine-grained sandstone and siltstone interbeds with mudstone a thin to moderately thickness 30-50 cm, reddish to yellowish in colour, massive and laminated and ripples, mud intraclast, calcareous silt and mud. Vertical and horizontal bioturbation are common in this facies (facies Fm and Fl) of Miall (1977) and Rust (1978). They differ from each other in their content of carbonaceous material (Figs. 3.7 and 3.8).

Interpretation. This facies is interpreted to represent crevasse splay deposits that occur throughout sequence of ASF at the Wadi Ghan section, and are interpreted to have been deposited in lobes adjacent to floodplain. They are typically well sorted, fine to very fine-grain laminated sandstones. These are locally rich in claystone and bioturbation. Sandstones display ripple cross-laminations and syn-sedimentary deformation. Crevasse splay deposits typically range in thickness from 0.2 m to 1m.

Facies 3b. Mudstone facies (Fsc) (Figs. 3.7 and 3.8)

Description. This facies (Fsc) consists of mudstone reddish in colour, massive, up to 10 m in thickness and up to 200 m in laterally in this section (Figs. 3.7 and 3.8) The main features of this lithofacies, which distinguishes it from Fl,

Interpretation. This facies is interpreted as floodplain deposit somewhat more distal relative to clastic sources such as nearby fluvial channels.

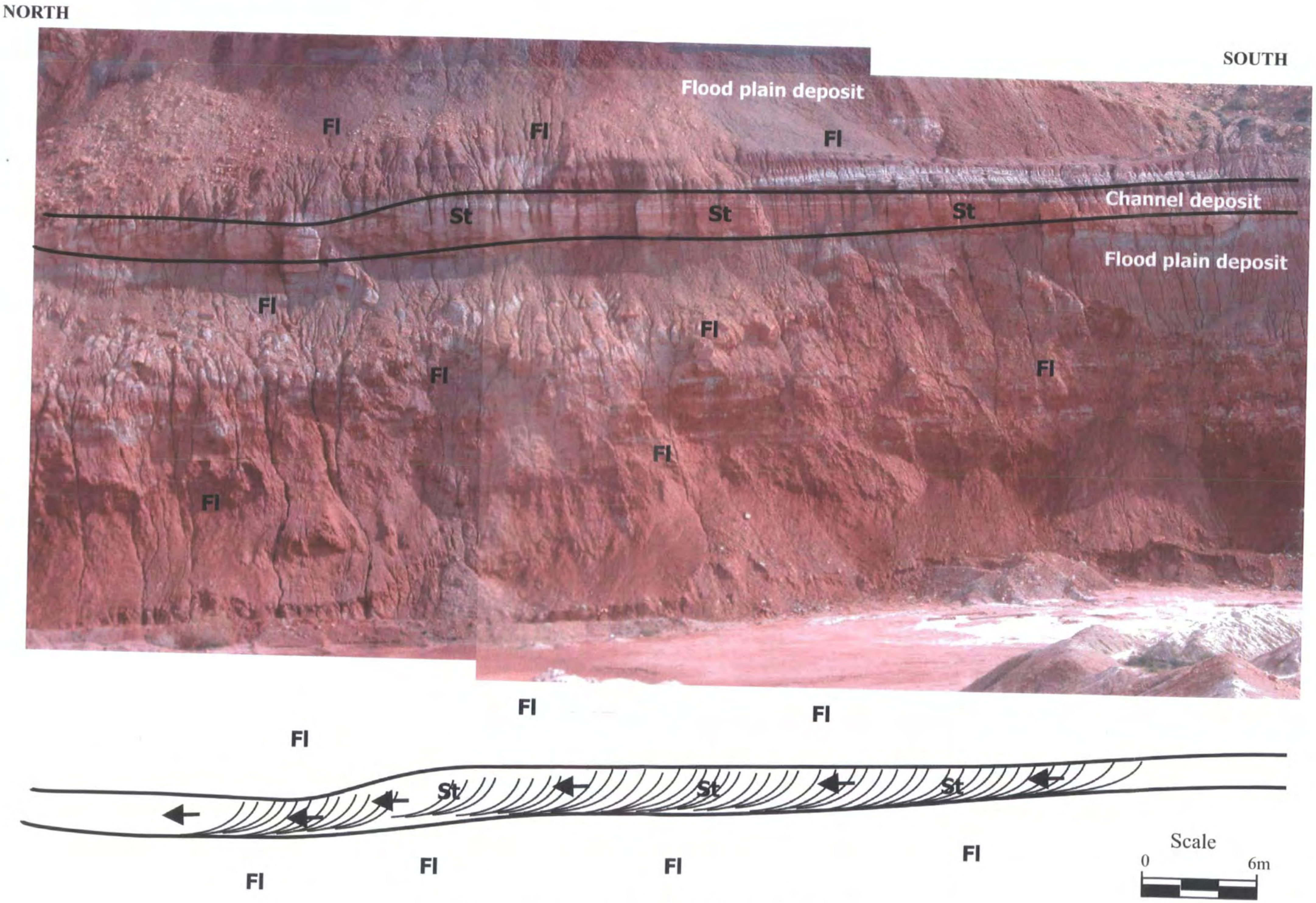
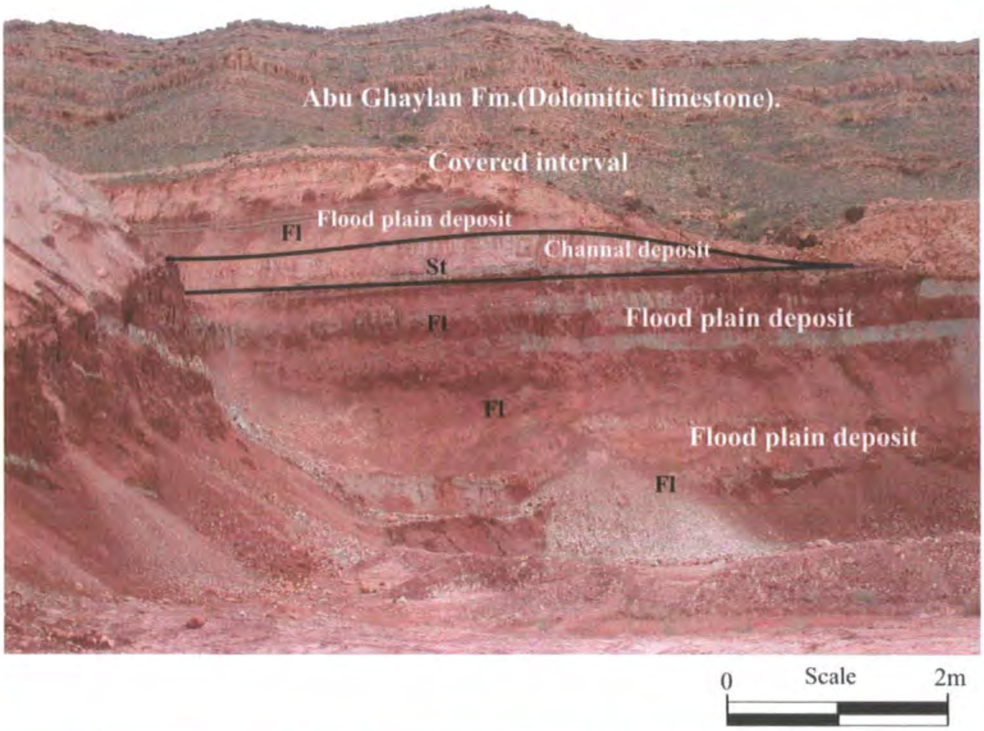


Fig. 3.7. Facies association 3; Flood plain (FP), facies 3b



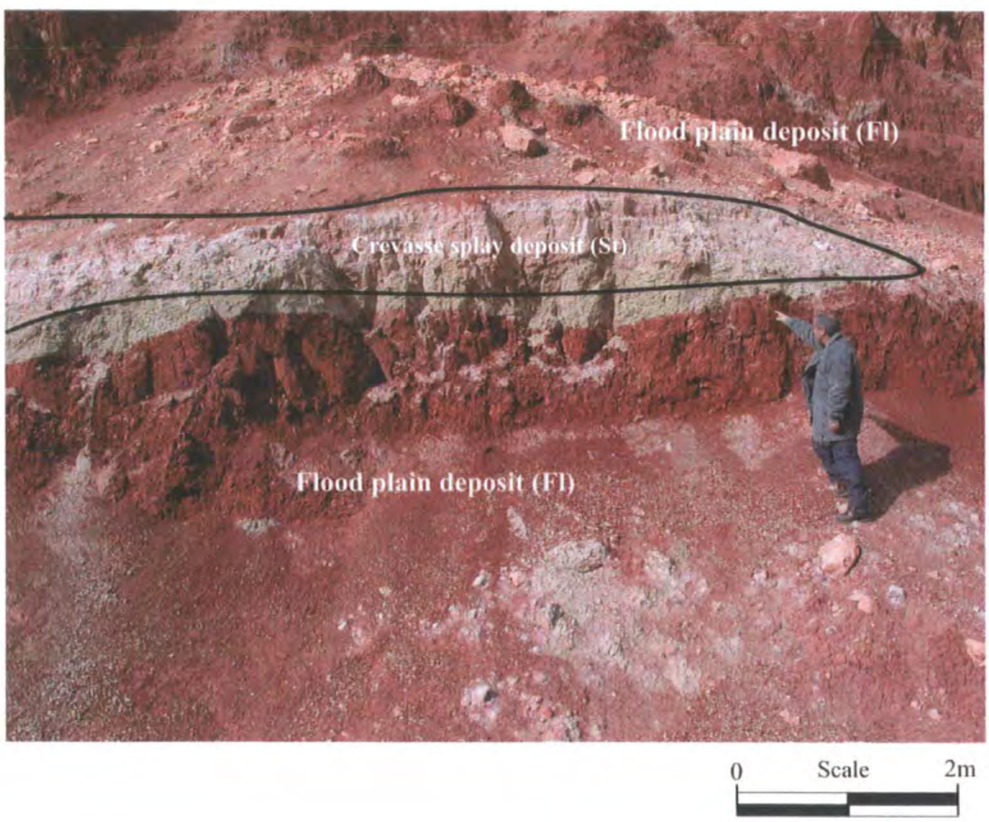


Fig. 3.8. Facies association 3: Flood plain (FP), facies 3a, 3b

Facies Association 4: Transitional zone (marine to continental lithofacies)

This facies is located in Abu Rashada road in the west of Gharyan dome (Fig. 3.1). Its consists of interlayered dolomitic limestone and fine sandstone with claystone and common horizontal burrows description (Upper Al Aziziyah Formation and base of ASF) about 21m thick (Fig. 3.9).

Facies 4a. Wavy ripples (Sr) (Fig. 3.10)

Description. This facies consists of interlaminations of mudstone, siltstone, and very fine-grained sandstone.

The thickness of particular sharp-based bodies varies from 0.1 m to more than 1 m. They pass upward into fine-grained, wavy ripple bedded sandstone, and interlaminated dolomite limestone (Fig. 3.10).

Interpretation. Sediments of this facies were deposited mainly as fine grained, intercalated clay with silt. They most probably represent a transitional zone from marine to continental.

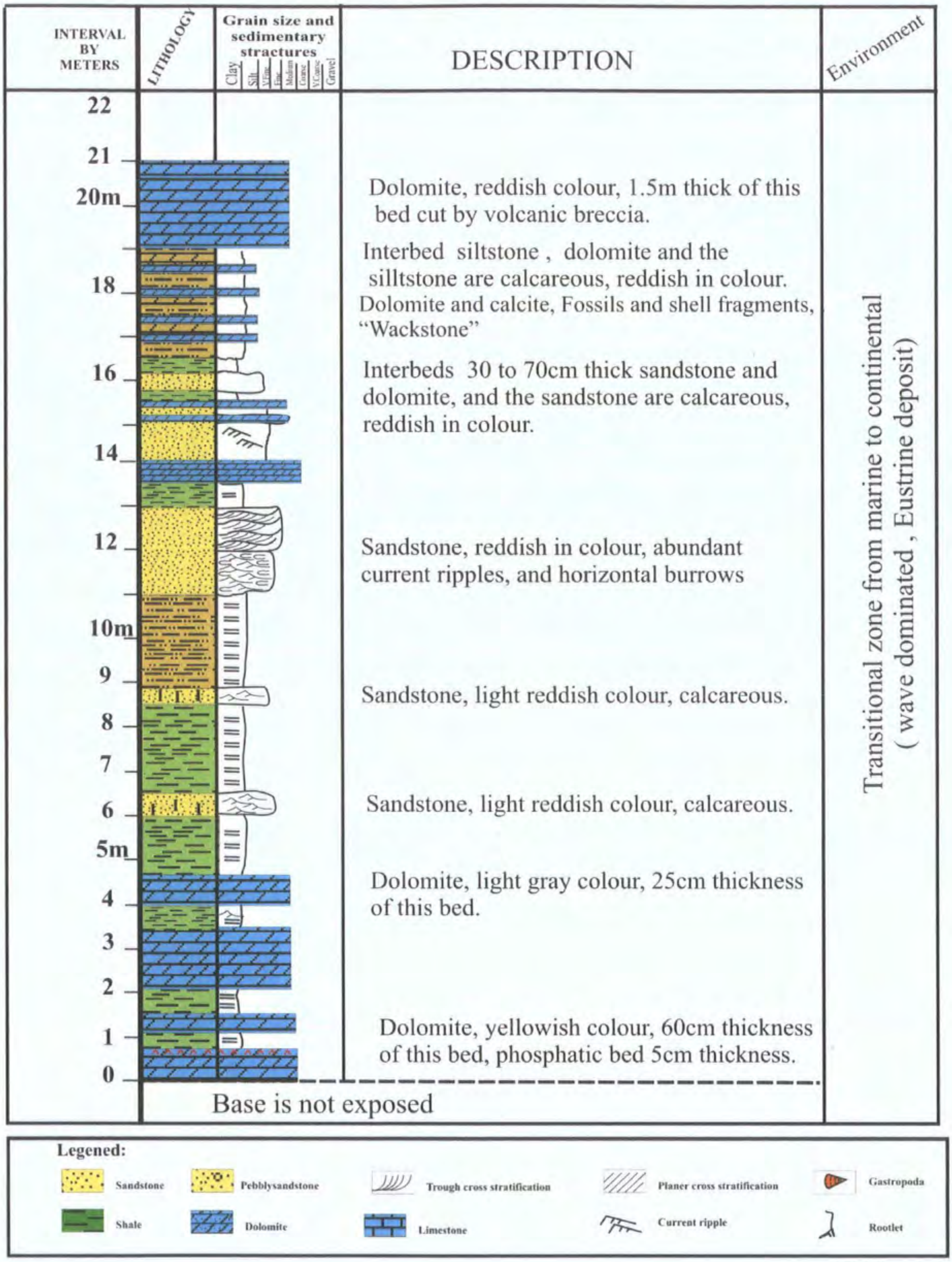


Fig. 3. 9. Log of Abu Rashada section showing the transitional zone.

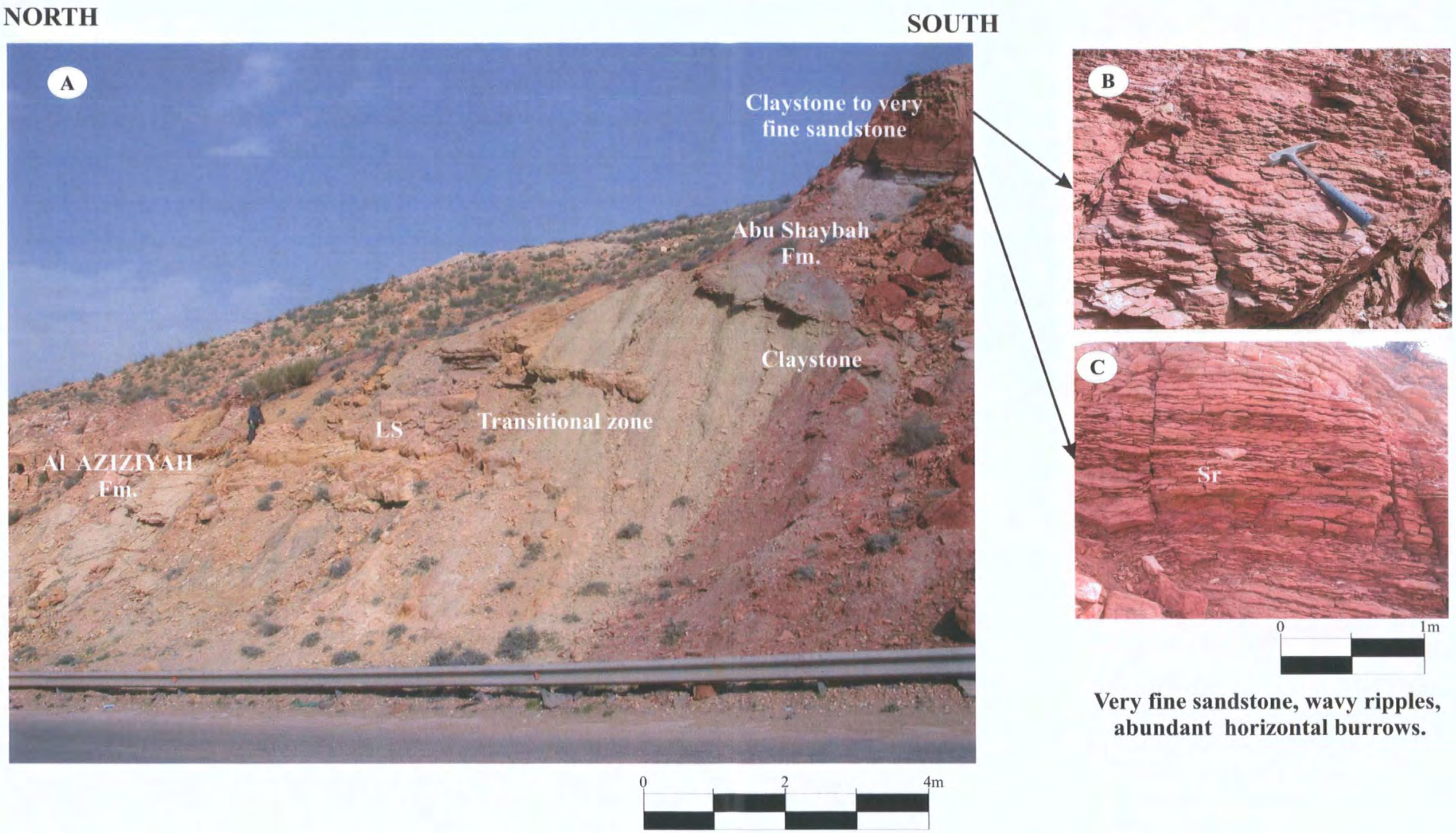


Fig. 3.10. Facies association 4: (A) Showing transitional marine top of Al Aziziyah and base of Abu Shaybah Formations. light greenish to reddish colour mudstone, (B and C) wavy ripples.

3.4. Facies model of the depositional system

Wadi Ghan zone area is one of the most spectacular in the region, largely since many of its component faults crop out in the Jabal around Wadi Ghan. The component faults display two dominates trends, NW-SE and WNW-ESE and commonly abut against each other to produce a rhomboidal to anastomosing framework of NE-SW faults. This zone is approximately 25 km wide and can be followed from the southeast, where it appears from under a cover of Tertiary lava flows, for a distance of 50 km northwest wards towards Al-Aziziyah south of which it merges with the Al Aziziyah Fault zone. In this region, several major NW trending faults of the Wadi Ghan zone abut against one on another defining a highly characteristic rhomboid pattern.

Fig. 3.11 illustrates the importance for not only analysing the vertical succession, but also the lateral extent of the individual architectural elements. Deposition occurred in a range of fluvial subenvironments.

The basal facies is conglomerate, sandstone (Facies 1a-1c) deposited in a fluvial braided system (FB1and 3)) that changed vertically and laterally from massive to horizontal cross-stratification (Facies 1a) and associated channelised conglomerate (Facies 1b and 1c).

Deposition of facies association 2a-2c was characterised by broad channel complexes of fluvial braided, sand sheets and numerous mid-channel bars with accretionary foresets (Fig. 3.11). The increase in grain-size, deeper channels and number of macro forms in this middle and upper part profile suggest that the fluvial system had a greater stream power and an increased sediment supply. Trough and small scale planar cross-bedding are interpreted as dune fields migrating downstream, which are commonly overlain by clearly defined laterally extensive sheet sands. The large–scale lateral and downstream accretionary foresets are the result of marginal and mid-channel bars. The overbank fine grained sandstones and siltstones are interpreted as having been deposited during flood stages and the immediately following waning stage, for each of the facies association units (FM 2 and).

Deposition of facies association 3 is characterised by mudstone and siltstone deposited on a flood plain. This facies is important in subsurface reservoirs for correlation of well logs.

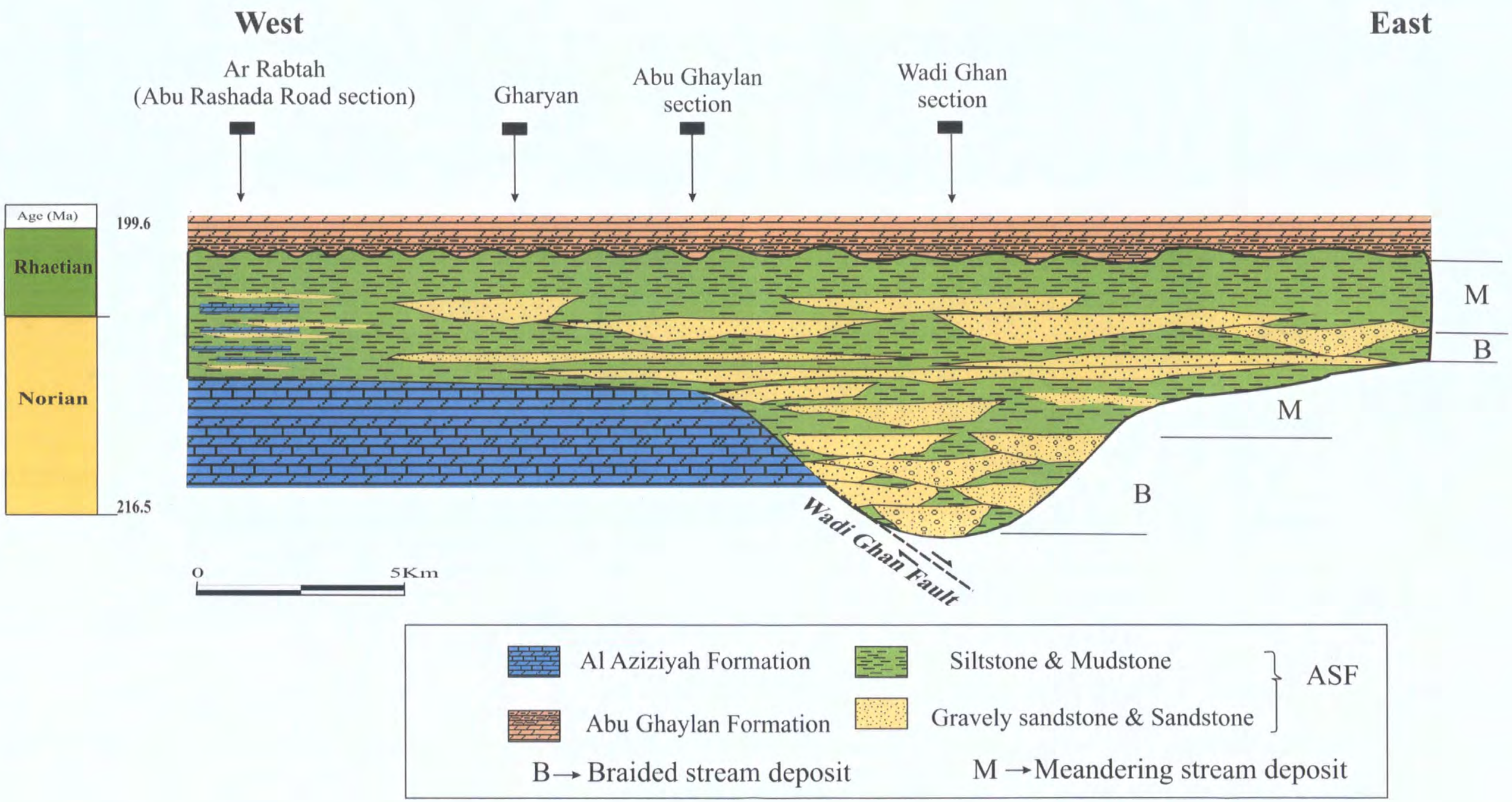


Fig. 3. 11. Depositional model showing lateral and vertical variation of ASF along Jabal Nafusah, NW Libya.

3.4.1 Four phase interpretation of Early – Late Triassic (249.7 -199.6 Ma)

The boundary between Aziziyah Limestone (Shallow marine to continental and Abu Shaybah is an unconformity and characterised by deep fluvial erosion, with the development of an incised valley system (Rubino et al., 2002). The upper Al Aziziyah Formation in the Wadi Ghan section consists of several meters of shallowing - upward cycles showing storm and wave dominated features.

First stage, **Early Triassic - Middle Triassic (Marine to Non Marine)**

The first stage of deposition of the ASF in Wadi Ghan section in incised valleys began after the end of marine sequence or surface erosion top of Al Aziziyah Formation, Late Triassic (Carnian 228 -216.5 Ma). The lithostratigraphic formation boundary between the Al Aziziyah and Abu Shaybah Formations is marked by a major unconformity and hence it is also a chronostratigraphic boundary.

The fluvial braided consists of conglomeratic sandstone with scattered intraclast at the base of trough and planar cross-bedding.

Second stage, Middle - Late Triassic (From Braided to Meandering system)

The second stage started by deposition of the extensive sheets of sandstone complex channel of braided river (CH) (facies association 1a, b and c) in the incised valley , this channel showed that braided system deposits consists of blocky sand units made up of superimposed 1 to 10 m thick megaripples deposited by low stand system tract and changed gradually to meandering system defined by thinning and decrease in grain size and climbing ripples (Facies association 2a,b and c) and the channels in this system is clear lateral accretion by trough and planer cross-stratification (St, Sp and Sr), intercalation by 0.5 - 5 m thickness of mudstone and siltstone (Floodplain and crevasse splay deposits) facies association 3 . This phase is braided river sit on major unconformity and characterized by fluvial erosion, with the development of an incised valley fill system.

Third stage, Middle - Late Triassic (From Meandering to Braided system)

A change from meandering to braided river patterns can be triggered by a number of causes. At a given discharge, braided channels occur on steeper slopes than do meandering channels. On the basis of this relation, a change from meandering to braided morphology can be inferred to reflect an increased channel slope and increase in clastic sediments sandstone body as channels thick channels of trough cross stratification 3-5 m, The vertical evolution of the fluvial system in a single sequence records a strong decrease of the slope angle of the fluvial profile. The sequence is thought to record the composite effected by tectonic and climate.

Fourth stage, Late Triassic – Early Jurassic (Continental to marine)

This stage transitional phase from continental (meandering stream deposit) to marine deposits of the Abu Ghaylan Formation, 20 cm thickness of limestone algae mat at the contact of Wadi Ghan section with abundant marine fauna and contorted interbeds of carbonate and marl with ripple marks, mud cracks, solution collapse breccias and tepee structures . The contorted bedding has been related to syn-sedimentary tectonic influences (Hammuda *et al.*, 2000).

3.5. Sequence Architecture and Stream Equilibrium Profile Fluctuation in the ASF

Fig.3.13. shows extensive alluvial unconformity surfaces and their correlative conformities. Deposition of the alluvial suite commences with amalgamation of channel sand bodies due to a lack of accommodation space causing local lateral migration or avulsion of the channel units, floodplain reworking, abandonment and preservation of channel sands before lateral avulsion or migration of the channel system re-occurs.

Sequence boundaries in the non-marine realm are caused by a lowering of the alluvial systems Stream Equilibrium Profile (SEP) with respect to the actual stream profile. The extent to which the SEP is lowered relative to the actual stream profile, and the lag time until a response occurs governs whether a confined or unconfined sequence will develop. As in coastal and marine sequence stratigraphy, two types of sequence boundary can be differentiated (Fig. 3.14). The" confined amalgamated channel unit" is the geomorphologic equivalent of the incised valley system and Type I sequence boundary of Van Wagoner *et al.*, (1988). The "unconfined amalgamated channel unit" is the geomorphologic equivalent of the Type II sequence boundary. The terms Type I and type II sequence boundaries are not used here in this model of continental sequence stratigraphy as they imply a dominant relative sea-level control on the depositional system.

Similar types of strata bounding surfaces occur within sequences in hinterland regions (Mitchum et al., 1977). Confined continental sequences are bounded below by an erosional valley surface (caused by an SEP Low) and above by a confined or unconfined sequence boundary. Unconfined continental sequences are bounded below by an erosional surface (again caused by an SEP Low) which follows the bases of the lowermost erosive channel sand units which constitute the amalgamated channel sand sheet resulting from a stream equilibrium profile relative Lowstand. An unconfined sequence is bounded above by either a confined or unconfined continental sequence boundary. Unless a confining incised valley can be demonstrated, amalgamated channel sands should be described as being unconfined, SEP Lows can occur without the development of incised valleys.

Wright and Marriott (1993) and Shanley and McCabe (1991, 1994) have subdivided continental sequences into Lowstand (LST), transgressive (TST), and Highstand (HST) systems tracts to temporally equate environments. However use of terminology such as transgressive systems tract in environments where influence of relative transgression on the coastal plain has a low probability of being physically expressed further upstream. (i.e., over most of the alluvial plain), can create confusion in correlation of strata surfaces (e.g., the coastal type I; Van Wagoner et al., 1988) and regional erosional surfaces that develop entirely within the subaerial realm. Due to the differing importance's of allocyclic controls in different reaches of the fluvial system, a sequence boundary developed as a result of a relative sea level fall at the coast. However, the two different sequence boundaries may tie together at some point in the subaerial depositional basin, even though they are not chronostratigraphically equivalent.

Fig. 3.14 illustrates the continental phases of deposition and erosion which generate the subaerial equivalent of systems tracts, developing as a result of fluctuation of the stream equilibrium profile. The amalgamated channel sands within an incised valley constitute the initial deposits of the Confined Stream Equilibrium Profile Low (CSEPL) of Fig. 3.14. The fine grained sediments of this unit have poor preservation potential due to lateral switching of, reworking by, and preservation of channel sands. A CSEPL, essentially generates a continental sequence boundary, or incised valley surface. On the alluvial plain it is unlikely that this will be caused by a drop in relative sea level. Soil horizons may develop (Bown and Kraus, 1987; Wright and Marriott, 1993) on well drained river terraces produced by a low stand event, and these could represent some of the correlative conformities corresponding to the hiatal sequence boundary. The sediments of the CSEPL are comprised mainly of amalgamated channel sand bodies deposited within the confining incised valley system. These sandy units are frequently exploited for mineral and hydrocarbon resources. As aggradation continues, so the incised valley fills.

As the Stream Equilibrium Profile fluctuation cycle moves on into the Stream Equilibrium Profile High (SEPH) the rate of aggradation increases, causing channel sand bodies to become more isolated within better preserved fine grained sediments of the floodplain (Figs. 3.13 & 3.14). Very early and very late SEPH floodplain deposits, where rates of sedimentation (aggradation) are quite low, may be dominated by soil development. However, when aggradation rates increase, development of thick mature soil horizons will be inhibited, and floodplain fines will be preserved without significant pedification. Posamentier and Vail (1988), Posamentier et al., 1988, Shanley and McCabe 1994 and the present conceptual model all suggest that alluvial flooding surfaces reflect rapid aggradation. Alluvial flooding surfaces are regionally extensive subaerial deposited complexes of fine grained floodplain deposits, which exhibit a higher degree of preservation of original sedimentary structure (i.e., have a low degree of pedification). This contradicts the model of Wright and Marriott (1993, Fig. 1) which suggests that amalgamated channel sheet sands are generated by a Stream Equilibrium Profile High.

Unconfined Stream Equilibrium Profile Lows will develop upon lowering of the rate of aggradation after the relative SEPH, causing a higher degree of channel amalgamation and reworking of any floodplain fines or soil complexes that have developed. Given enough times a regionally extensive amalgamated channel sheet sand will develop due to low rates of elevation change. If the stream profile remains above the stream equilibrium profile then the system will continue to degrade, incising into its previous deposits, possibly generating the deposits of the Confined Stream Equilibrium Profile Low as previously discussed.

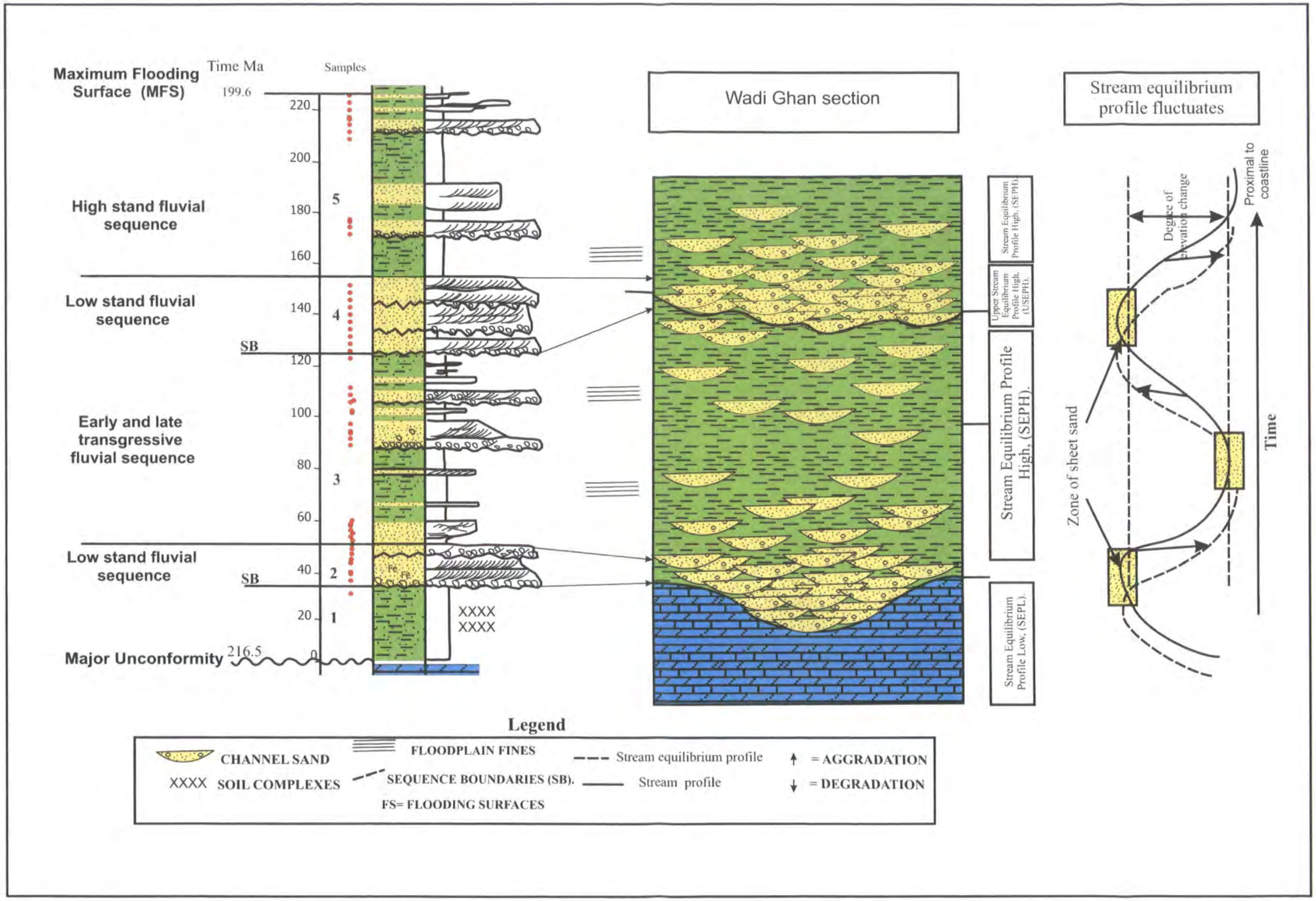
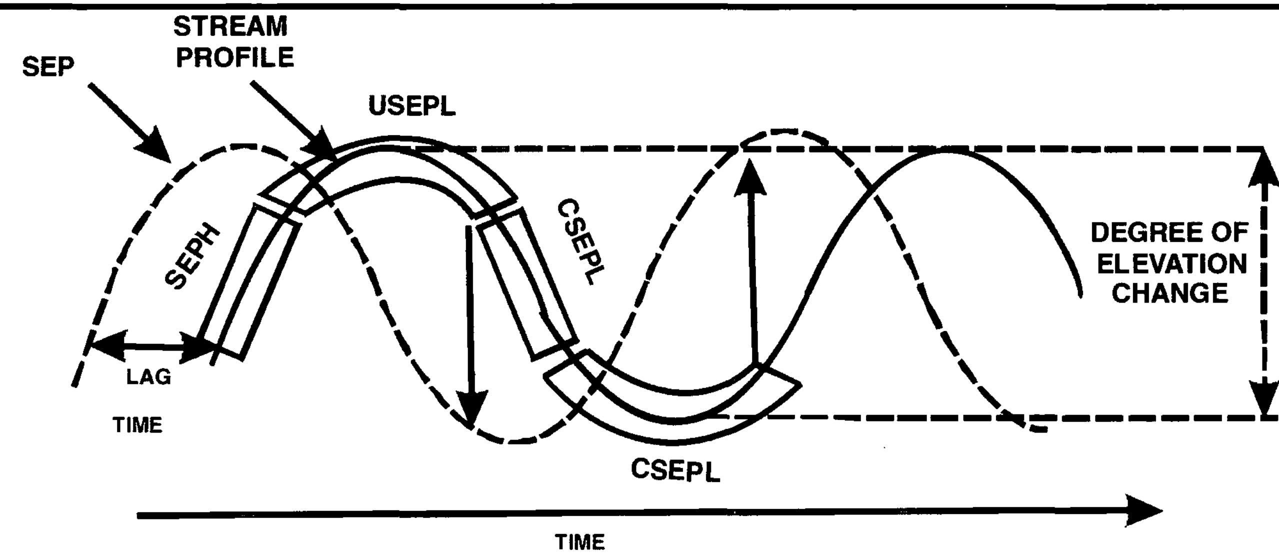


Fig. 3.13. Illustration of gross sequence architecture and continental systems tracts of Abu Shaybah Formation, (ASF) in Wadi Ghan section, which develop with stream equilibrium profile elevation fluctuation, resulting from alteration of allocyclic controlling mechanisms.



Keys

SEP - Stream Equilibrium Profile

SEPH - Stream Equilibrium Profile High

CSEPL - Confined Stream Equilibrium Profile Low

USEPL - Unconfined Stream Equilibrium Profile Low

Fig. 3.14. Continental system tracts and their time to elevation with respect of fluctuation of the stream profile and the stream equilibrium profile.

3.6. Discussion

3.6.1. Climatic controls on fluvial sedimentation

Regional climatic variations arise as a function of plate positions and how they change through time. Climate directly controls aridity, humidity and rainfall patterns and as a direct result perhaps the fluvial environment is one of the most sensitive indicators of climate change in the rock record (Frostick & Jones 2002; Jones & Frostick 2008). Climate has a direct impact on erosion, drainage patterns and deposition within a sedimentary basin. Any changes in climate can have a marked and devastating change for the whole of a river system (Blum and Tornqivst, 2000).

Studies of climate change controlling fluvial deposition in the Pre-Quaternary are rather sparse in the geological literature compared to the vast array of literature on sea-level and base level changes following the introduction of sequence stratigraphic concepts. However, by far the most widely used climate indicator are palaeosols in floodplain successions as outlined by several workers over recent years (e.g. Retallack, 2001; Kraus, 1999; Wright and Marriott 1993). Instead the discussion below focuses firstly on where workers have specifically interpreted depositional sequences in response to climate change and secondly how climate variations can be determined in the fluvial packages of the Late Triassic Abu Shaybah Formation (ASF) in this study.

Many workers have inferred climate change through the use of a variety of proxies or transition from fluvial to other strata. These are highlighted below:

1. Changing from braidplain environments then to a meandering stream dominated setting that pass laterally environments, were attributed to shifts towards from arid and then more humid conditions. A number of studies also interpret transitions between fluvial and other strata due to climate changes over time (Cojan 1993; Yang & Nio, 1993; Olsen & Larsen 1993; Clemmensen et al., 1994; Dubiel *et al.*, 1996 , Smith 1994 ; Huisink, 1999)
2. Difference thickness of floodplain and channels, Olsen (1990, 1994) identified two scales of Cyclicity in the thickness of channel sand bodies.
3. Difference in sediment supply stages (aggradation and degradation) Smith, (1994) suggested that changes in channel sandbody geometry, sand-to-mud ratios, sediment accumulation rates and types of palaeosols in Arizona (USA) occurred during a tectonically quiescent period and might instead reflect climate changes.
4. The relationship between stream discharge and sediment flux under a range of climate regimes (Lane 1955; Bull 1991; Demko *et al.*, 2002; Meadows 2006;

Jones & Frostick, 2008), in which net fluvial aggradation or degradation is strongly influenced by the relationship between bedload and stream discharge.

It is very easy in this study to ascribe the practise as outlined above in 1 to 3 as not only is the ASF located between two marine successions (Table 2.1), but also the ASF changed gradually from a braided river as part of an incised valley fill to a meandering stream with isolated channel sandstones enclosed by siltstone and mudstone of floodplain deposits. Such a sequence is similar to that predicted by Shanley and McCabe (1994). It is obvious that climate has played a role in the behaviour of the ASF fluvial systems at a sequence scale, but more important is how climatic changes can be recognised at the outcrop and used in correlation of fluvial facies over 10's kms.

According to the work of Meadows 2006 (Demko *et al.*, 2002 *therein*) they have modelled the effects of climatic change on patterns of fluvial discharge and sediment flux under a range of differing climatic regimes. In considering climate change Demko et al., (2002) looked at a range of climatic regimes from wet and tropical monsoonal climates through to dry seasonal and desert climates, the types of vegetation associated with these and the relative proportions of chemical and mechanical weathering. In doing so it was concluded that the greatest sediment flux is associated with wet and tropical monsoonal climates, but that the vast majority of this comprises clay-and silt grade material liberated through chemical weathering. With progressively drier climates the equilibrium sediment flux decreases but becomes increasingly dominated by sand-and gravel-grade material liberated by mechanical erosion.

If such a methodology is applied to the ASF then it can clearly be demonstrated that a cyclicity exists within the fluvial succession from being wet seasonal through dry seasonal to relatively dry and with an increasing proportion of sand grade material the finer-grained sediment becomes less abundant (Fig. 3.15).

The climate change model therefore provides a means by which the maximum sediment flux and the maximum stream discharge are not in equilibrium. Rather, there are episodes during which the sediment flux increases dramatically, generating conditions favourable for rapid net deposition and aggradation, and episodes during which discharge increases without a concomitant increase in bedload (Fig. 3.15). At such times, the river would be likely to undergo net degradation, reworking previously deposited sediment and incising its floodplain. Demko *et al.*, (2002) proposed an idealized stratal architecture resulting from this imbalance that illustrates the effect of alternating phases dominated by degradation and then aggradation. Interestingly, this also generates incised valley systems filled with stacked amalgamated fluvial channel deposits (Facies association 1-braided stream) overlain by isolated fluvial channel sandstones enclosed within floodplain deposits (Facies association 2 & 3 – meandering and flood plain). In this case, it can be inferred that incision occurs during the most pluvial part of the climatic cycle, that sediment flux and discharge become effectively balanced as the climate becomes arid and, presumably, during this phase the river remains trapped within the previously incised valley system. As the climate begins to become wetter, and stream discharge increases, the sediment pulse is released and the river aggrades rapidly, infilling the incised valley until it breaks out onto the wider floodplain. At this stage both the increase area of the floodplain and the declining sediment flux, especially the coarser –grained bedload component, increase the preservation potential of floodplain facies (facies association 3) and therefore render it more likely to generate isolated channels encased within those deposits.

The main climate control on drainage development is through discharge variability It is to be expected that lithofacies heterogeneities will increase with increasing seasonal or long–term variability in discharge, and that the frequency of the minor bounding surface will increase (i.e. second and third order surfaces; Miall, 1996). The lithofacies and change in fluvial style in the ASF during the Triassic undoubtedly can be explained by changes in discharge sediment ratios (or sediment flux) and correspondingly discharge from waning flood events. The succession of facies described here is known to span the time of gradual continental break-up of Pangaea and a change from icehouse to greenhouse conditions with a trend towards more temperature-humid conditions. The gradual change in climate coincides with changes in character of the fluvial systems from dominant coarse-grained braided rivers to meandering streams. However, river react to a number of stimuli, and it is difficult to determine whether a given change in character reflects tectonic or climatic influences, or both (Frostick & Reid, 1989; Frostick & Jones, 2002). The evidence the ASF is affected by climate control is contained in that area red beds are typical of semi arid regions and many form through the release of iron from mafic minerals during early burial and its precipitation around grains as a hydrated iron oxide, which ages to haematite (Tucker, 2001).

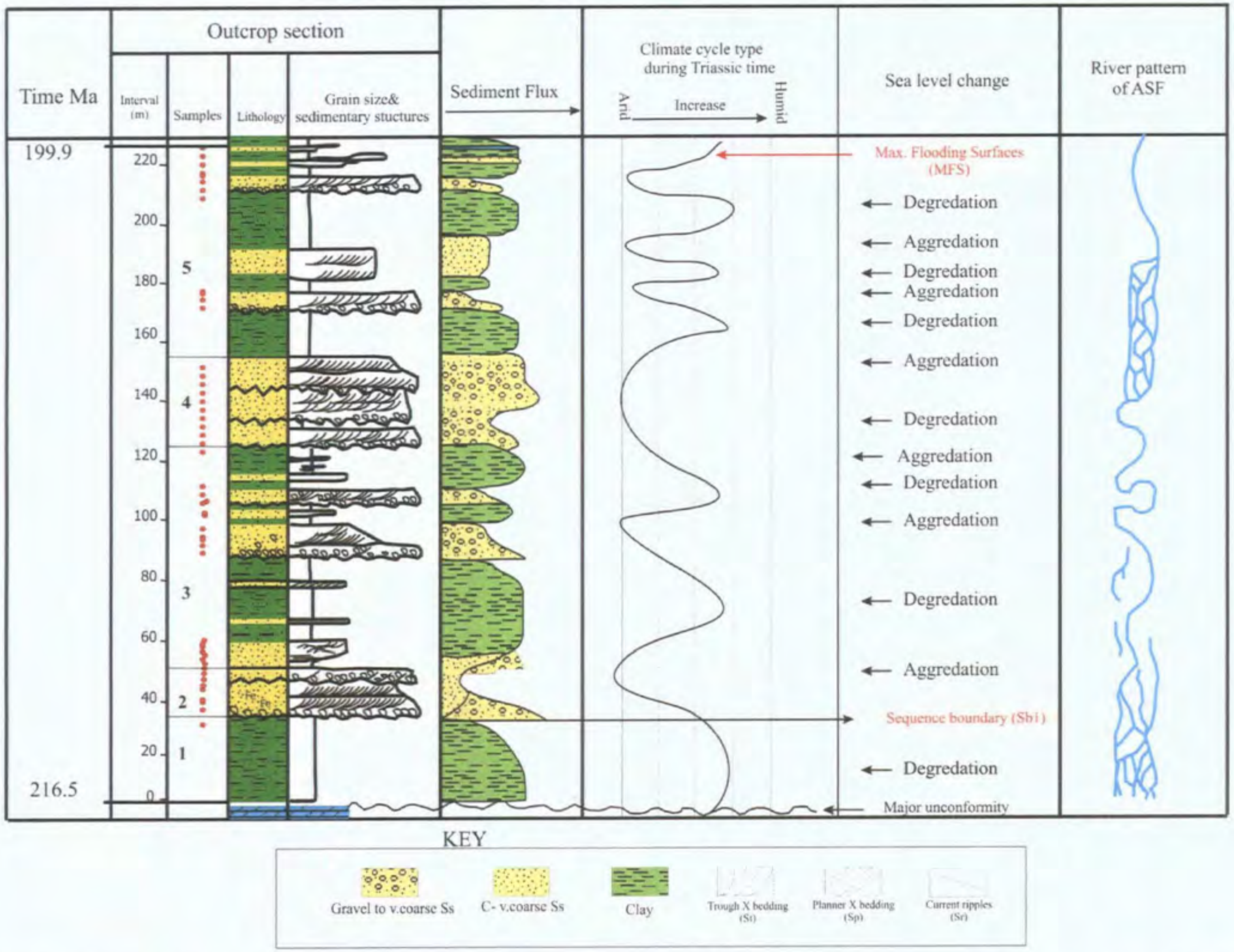


Fig. 3.15. Relationship between climatically mediated stream discharge and sediment flux under a range of climatic regimes during Late Triassic for the Wadi Ghan section, NW Libya (concept adapted from Demko *et al*., 2002; Meadow, 2006).

When interpreting fluvial successions climate alone cannot reliably be used due to the complex interactions and feedback loops associated with tectonism (e.g. Frostick, 1993; Frostick & Jones 2002). For the purpose of this study the tectonics of the Jifarah basin has been well constrained as outlined in Chapter 2, but will be further developed in the next section.

3.6.2. Tectonic control on fluvial successions

Background

Rivers are sensitive indicators of tectonic subsidence, uplift and tilting. In tectonically active areas, and especially during synrift phases of basin evolution, tectonics can be recognised as an important control on gradients, water supply, stream power, shear stress, long profiles, terrace patterns and channel sinuosity (e.g. Seeber & Gortniz, 1983; Ouchi, 1985; Rasanen *et al.*, 1987; Jones 2002; Frostick & Jones 2002). Frequently in the literature tectonics is ascribed as one of the main controls on fluvial architecture, geometry of channel sand bodies, spatial distribution of fluvial systems and direct control on sediment flux routing into sedimentary basins (e.g. Gawthorpe & Leeder, 2000). However, process-based models of alluvial stratigraphy have been developed that consider the effects of compaction, tectonic tilting of the floodplain and variation in the aggradation rate with distance from channel belt (Bridge & Leeder, 1979; Mackey & Bridge, 1995). Such predictive process-based models are widely used in understanding fluvial systems and routinely incorporated into reservoir models for petroleum exploration. Mackey and Bridge (1995, p. 28) identified that even with advances in the models several components are still lacking a complete understanding of ancient fluvial behaviour (e.g. sediment flux, channel diversion occurrence, fluid flow, creation of levees, lakes and crevasse splays). There have been several attempts at developing physical based three-dimensional models of fluvial deposition (reviewed by Kolterman & Gorelick 1996; North 1996; Paola 2000). However, none of these approaches has been able to successfully simulate three-dimensional fluvial architecture to a satisfactory level of detail.

3.6.3. The Jifarah basin and the ASF

The Jifarah region has previously been discussed in Chapter 2, but the importance of the tectonics of the region is further emphasised here. Uplift in the southern Jifarah region during the late Triassic accounts for the unconformity at the base of the Abu Shaybah Formation and the onset of continental conditions of sedimentation at this time. Both the NW-SE and E-W structural grains associated with the Caledonian and Hercynian structures in the Jifarah basin are of importance in their influence on later structural events (Mikbel, 1977; Goudarzi, 1980; Anketell & Ghellali, 1991). Faults within the region display a characteristics *en echelon* arrangement that are interpreted in terms of long-lived reidel shears and imbricate fan splays related to strike-slip fault movements on deep seated basement fractures (Anketell & Ghellali, 1991).

Due to the tectonic activity of the research area and many of the outcrops located near important strike-slip faults along the Jabal Nafusah escarpment (e.g. Coastal zone, the Al Aziziyah zone in the west and central Jifarah, the Tiji zone in the foothills of the Jabal Nafusah and, the Nalut zone exposed at the top of the escarpment; Figs. 1.7 and 2.9), it is tempting to speculate that the fluvial systems of the ASF were strongly controlled by tectonic activity and perhaps many of the coarse beds represent tectonically induced erosion and sediment flux from the hinterland catchments (e.g. Blair and McPherson 1994; Mack and Leeder, 1999; Viseras *et al.*, 2003).

However, from field observations and geological mapping it is evident that the Late Triassic fault activity is important in controlling the gross geometry of the fluvial succession and hence their drainage patterns for the ASF. Recent studies have highlighted the importance of tectonic grain and changes in fault orientations in controlling drainage patterns and hence outlet spacing into a depositional basin (e.g. Hovius, 1996; Jones et al., 2001; Jones, 2004). Furthermore, the physiographic changes to the landscape, induced by fault activity are likely to have far-reaching implications on fluvial style and flow directions.

The incised valleys fills of ASF are most likely to have formed through a combination of both erosional and tectonic forces (e.g. Stanley, *et al.*, 1994). This is reflected in the basal erosional sequence boundary of the ASF (Fig. 3.13). The amalgamated fluvial deposits that are confined within the incised valley were further controlled by localized faulted margins. Through the Late Triassic (c. 15.9 Myrs) the fluvial systems changed character and became less influenced by tectonism reflecting fewer multistory/amagalmated fluvial deposits and more isolated channel networks. Any changes in the river behaviour can be attributed to regional gradient changes of the basin floor or minor diversion by reactivation of basement strike slip faults.

Although it has been widely recognised that fault movement, fault geometries and subsidence create a topography and control on fluvial patterns and deposition, it is climate that exerts a more discrete control on sedimentation of fluvial sequences and their architecture. This is perhaps too easily overlooked in ancient basin settings, but as recognised in this study, climate is a more important control on fluvial sequences once a drainage pattern has been created.

This study has demonstrated that although tectonism can induce and control fluvial system development, it is climate that exerts a strong influence upon the nature of the fluvial sequences and how they interrelate to associated facies and can be correlated. This is easily overlooked in the need to identify tectonic events and interpret ancient basin history. Furthermore it is parameters related to climate (e.g. rainfall, stream power) that need to be further understood in ancient successions to improve process based models of fluvial deposition.

3.7. Summary

Lateral and vertical variations in the facies architecture of the fluvial strata of the ASF along Jabal Nafusah, NW Libya can be explained by consideration of allocyclic controls on sedimentation. Tectonism in the source area was probably important in determining the timing and location of the alluvial wedge. Differential subsidence in the depositional basins influenced the location of braided rivers and determined the degree of interconnectedness of sandstone bodies. A gradual climatic change produced change in the style of rivers and the geometry of their resulting facies. Changes in stratigraphic base level resulted from shrinkage and expansion of lacustrine systems downstream from the study area. These changes in base level are interpreted to have caused variations in the rate of creation of accommodation which resulted in a sharp base to the formation and an upward change from thin, highly amalgamated, sheet sandstone bodies to thick, isolated, sheet sandstone bodies. The causes of the fluctuations in lake size are beyond the scope of this study, but it is possible that they were related to climatic shifts or, because of their position on a broad coastal plain, to changes in sea level in the seaway to the north.

This study emphasizes the importance of taking a holistic approach to understanding controls on fluvial sedimentation. The role of both upstream and downstream controls on sedimentation needs to be considered in addition to subsidence and climate at the site of deposition. The climate change model therefore provides a means by which the maximum sediment flux and the maximum stream discharge are not coincident. Rather, there are episodes during which the sediment flux increases dramatically, generating conditions favourable for rapid net deposition and aggradations, and episodes during which discharge increases without a concomitant increase in bedload. At such times, the river would be likely to undergo net degradation, reworking previously deposited sediment and incising its floodplain. Demko et al. (2002) proposed an idealized strata architecture resulting from this imbalance that illustrates the effect of alternating phases dominated by degradation and than aggradations. Interestingly, this also generates incised valley systems filled with stacked amalgamated fluvial channel deposits overlain by isolated fluvial channel sandstones enclosed within floodplain deposits. In this case, it can be inferred that incision occurs during the most pluvial part of the climatic cycle, that sediment flux and discharge become effectively balanced as the climate becomes drier and, presumably, during this phase the river remains trapped within the previously incised valley system. As the climate begins to become wetter, and stream discharge increases, the sediment pulse is released and the river aggrades rapidly, infilling the incised valley until it breaks out onto the wider floodplain. At this stage, both the increased area of the floodplain and the declining sediment flux, especially the coarse grained bedload component, increase the preservation potential of floodplain facies (sandflats in the case of the ASF) and therefore render it more likely to generate isolated channels encased within those deposits.

Several important conclusions can be drawn from the architectural study of the ASF that has implications for the development of drainage networks in the Jifarah Basin.

1. The depositional model comprises of fluvial system by low sinuosity braided, meandering stream, floodplain and crevasse splay deposited.
2. The architecture of the braided fluvial deposits are dominated by channel –fill complexes (Gst & GSp), sands sheet, and associated simple and more complex bar types. The meandering stream sediments record several fining –up successions interpreted as point bars.
3. The braided system was deposited under a more humid climate with high rates of reworking base level is interpreted to have fallen markedly and meandering system was deposited under semi arid climatic conditions, stable fluvial activity and floodplain storage.

4. The geometry and architecture of the two sequences can be described simply as Sequence (1) of braided system, stacked multi-storey sand bodies, with a sheet like geometry procuring an amalgamated high net / gross (80%) sand body complex. Sequence (2) of meandering system, isolated channel-fill sand bodies with multi-storey / multi – lateral component, producing low net to gross (20%) sandstone ribbons, with a low connection potential.
5. Channel (CH). Consists of gravel bars and bed forms (GSp and Gst). Gravel bar and bed form thicknesses are up to 3 - 5 in most cases and widths are 50m to 150m.
6. Sand bed forms (SB) consist of St, Sp, Sr and SL with sharp base. SB has sheet-like geometry and is often eroded channels. SB is usually about 1.5 - 2 m thick. Widths vary between 150 and 200m. The average width/thickness ration of 15:1 within the whole studied area.
7. Paleocurrent indicators from the ASF show the flow to ward the west direction based on cross stratification measurements' in lithofacies Sr, St and Sp
8. Based on the work of Demko *et al.*, 2002 and Meadows 2006 for alternating episodes of incision and aggradation through climatic mediation of the relationship between fluvial discharge and sediment flux controlled by a changing climate. This is supported by recorded changes in thickness of floodplain facies association 2 & 3 in which the humid parts while the channel of sandstones facies association 1 in arid part.

Chapter 4

Use Of Petrography To Understand Climate & Tectonics

Chapter 4
Use of Petrography to Understand Climate & Tectonics

4.1 Introduction

Diagenetic and petrographic studies of fluvial sediments have long been undertaken to appreciate the provenance and burial history of sediments (Tucker, 2001). In recent years the use of diagenesis to better understand sequence stratigraphy has been explored. It was recently documented that diagenetic alterations such as calcite and dolomite cementation might be distributed coincident with sequence stratigraphic surfaces (e.g. Tucker et al., 1993; Taylor and Curtis, 1995; Morad *et al.*, 2000; Ketzer *et al.*, 2003). The synergy between sequence stratigraphy and diagenesis enables the prediction of spatial and temporal distribution of diagenetic alterations, and ultimately post-depositional evolution of reservoir, fluid flow through reservoirs, and diagenetic compartments and seals.

Our knowledge about sequence stratigraphy has broadened and many researchers have realized that it is not always applicable to apply in all sedimentary environments and perhaps fluvial settings have been the most difficult to resolve in terms of correlation (Shanley & McCabe 1995; Blum & Tornqvist, 2000). However, through the use of diagenesis useful information on the parameters controlling near-surface diagenesis can be attained, such as i) pore water chemistry (McKay *et al.*, 1995); ii) residence time of sediments under certain geochemical conditions (Taylor and Curtis, 1995); iii) detrital composition of extra- and intra-basinal grains (Zuffa *et al.*, 1995) and iv) presence and abundance of organic matter (Cross, 1988). Sequence stratigraphy cannot be used to assist in all areas of diagenetic analysis and is unable to assist with palaeoclimate change and tectonics of the basinal setting. This is unfortunate as palaeoclimate change is considered one of the more important controls on fluvial architecture and sedimentation patterns but will be discussed later in this chapter (Huisink, 1999; Blum and Tornqvist, 2000 ; Meadows, 2006).

This chapter provides a comprehensive petrographic and diagenetic study of the Abu Shaybah Formation (ASF). This is integrated with sequence stratigraphy in order to unravel and discuss the spatial and temporal distribution of diagenetic alterations in fluvial sandstones of the ASF.

4.2 Methodology

This study is based on a total of 52 thin sections collected from the study area (Fig. 3.1). The thin sections were analyzed using a point counter in order to determine the percentage of quartz, feldspar, rock fragments, heavy minerals, mica, kaolinite, chlorite, mud matrix, calcite, dolomite, hematite and type of porosity. The analysis was carried out using a Swift automatic point counter and mechanical stage. Three hundred grains were counted per slide for statistically reproducible results and then the results recalculated to 100%. The porosity measurements were estimated by standard modal analysis techniques (Pettijohn et al., 1987), and the porosity of ASF classified into three types for counting purposes: primary porosity (0 – 13.1% mean 3.3%), secondary porosity (0 – 13.6% mean 4.12%) and microporosity (0-3% mean 1.5%), all analysis data are summarized in appendix 1.

4.3 Compositional analysis

The modal composition of the ASF provides evidence for composition of the source and diagenetic events that affected the post depositional burial history of the sandstone. All the results from the petrographic analysis and percentages are tabulated in a summary sheet (Table 4.1).

Classification	**No. of thin section**	**Facies (Chapter 3)**
Quartzarenite	14	FB ,(CH) & FM, (SS)
Sublithernite	24	FB, (CH) & FM, (SS)
Litharenite	4	FB, (CH) & FM, (SS)
Subarkose	2	FM, (FPD) & FM, (SS)
Lithicwacky	3	FM, (SS)
Carbonate samples from the upper and lower contact Fms.	**5**	**3** samples from lower contact. **2** samples from upper contact.
Total samples	**52**	

Table 4.1. The detrital particles in siliciclastics rocks in the ASF in Wadi Ghan section, NW Libya. All samples are classified by using Pettijohn *et al.*, (1987), which has seven divisions, (Figs. 4.1a, b).

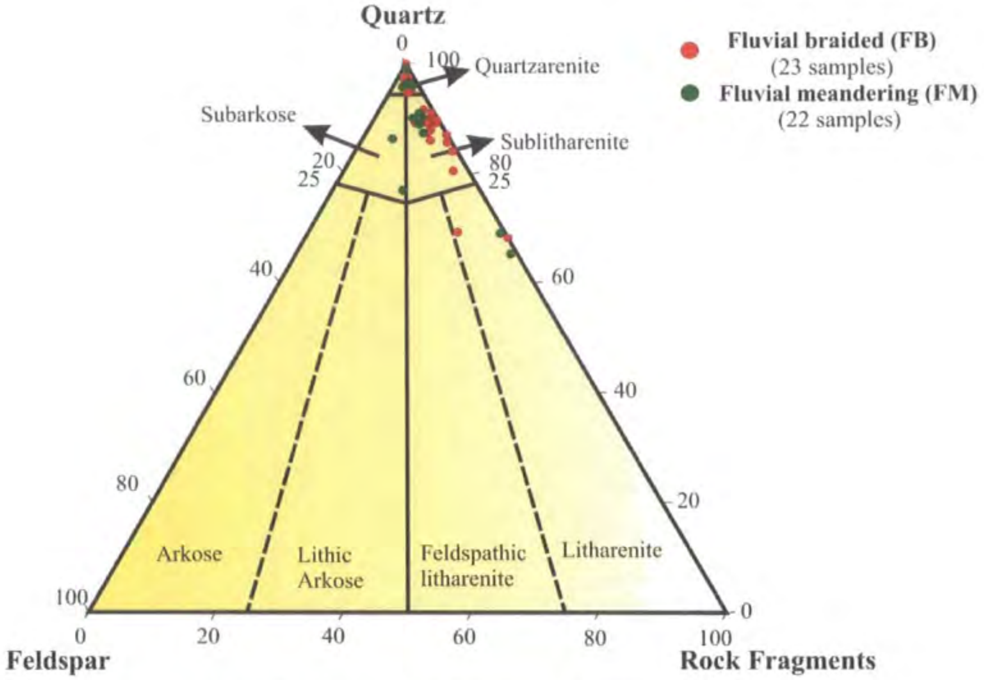


Fig .4.1a. Detrital composition of the Abu Shaybah Formation,(Wadi Ghan section, fig. 3.1) plotted on a Pettijohn et al., (1974) classification diagram.

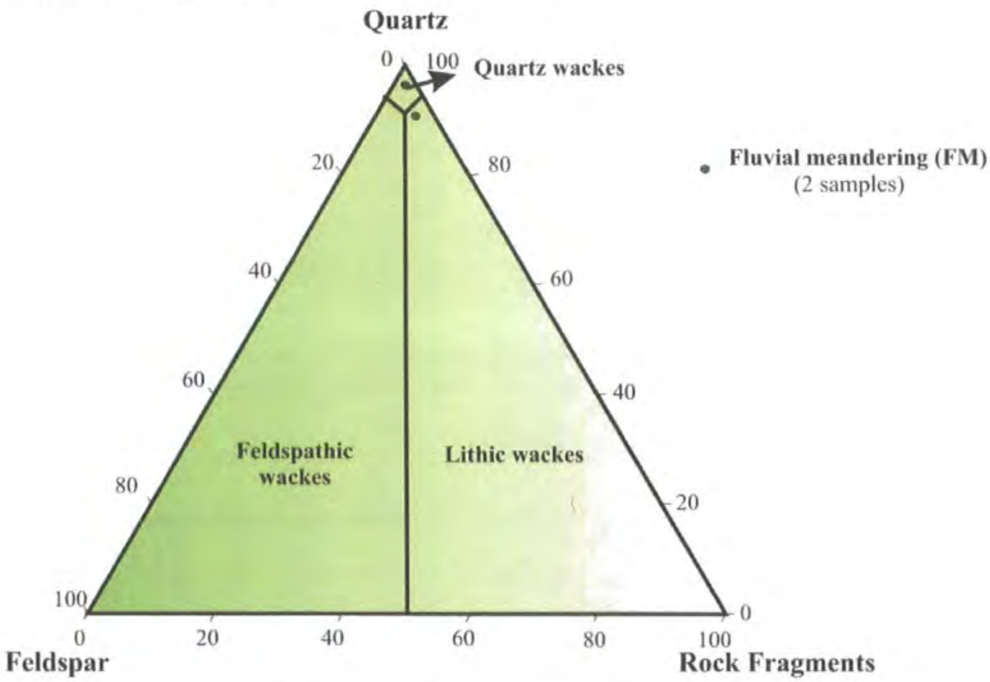


Fig . 4.1b . Detrital composition of the Abu Shaybah Formation,(Wadi Ghan section, fig. 3.1) plotted on classification diagram .

The sandstones in the ASF are composed mostly of quartz with a medial average of about 90%. The grains are predominantly monocrystalline quartz grains with subordinate polycrystalline quartz grains. Secondary authigenic quartz overgrowths, defined by inclusions and dust rims commonly occur. Most quartz grains are rounded to subrounded and feldspar is the second most important minerals in the ASF. The feldspar content varies from 0 up to 8 %. Feldspars in the ASF are mainly plagioclase and microcline, most of which have been partially or completely dissolved during diagenesis; other crystals are relatively unaltered. Rock fragments are up to 25 % from sedimentary rock fragments (Fig. 4.2). The presence of rock fragments provides some clues to source rock composition. Mica is a minor constituent of ASF. Muscovite is the most common mica seen in thin section (Tr – 6 % mean 3 %). Clay minerals within the sandstone are represented by vermicular Kaolinite, usually as a blocky filling in the pores between quartz grains, and chlorite as pore lining (Fig. 4.2) which are present in amounts up to 25 %. Kaolinite is the secondary mineral formed by weathering and /or diagenetic alteration of feldspar. Heavy minerals can provide evidence as to the composition of the source rocks, and they have proved useful in provenance studies (Tucker, 1991). Heavy minerals are present in small amounts of up to 1.5% in the ASF and are dominated by tourmaline and zircon.

The study of porosity is important for reservoir rock assessment. Porosity in the ASF results mainly develops from dissolution of the detrital grains and authigenic phases. During the mineralogical analysis increases in the porosity of the ASF was noted (Fig. 4.3), porosity values ranged from 0 % up to 23% (Figs. 4.2 & 4.3). However, no analysis of permeability of these samples has been attempted in this study because I haven't plug samples for analysis.

A. Macropores (Fig. 4.2)

Within these sandstones, point counted visible macroporosity varies from 0 to 23.9%. All these sandstones poor porosity, (mean 7.2 %), meandering sandstones, (mean 7.5 %), braided sandstones (mean 7 %) and the following macrospores types occur:

- **Intergranular macropores (0 – 13.1% mean 3.3%)** – It is very rare and the remnant intergranular pore spaces which have survived compaction and cementation.
- **Secondary dissolution macropores (0 – 13.6% mean 4.12%)** – principally comprised of pore space formed by both totally and partially dissoluted of feldspar grains and the breakdown of unstable lithic fragments.

B. Micropores (0 – 3% mean 1.5%)

Microporisty occurs as pores within kaolinite clay minerals.

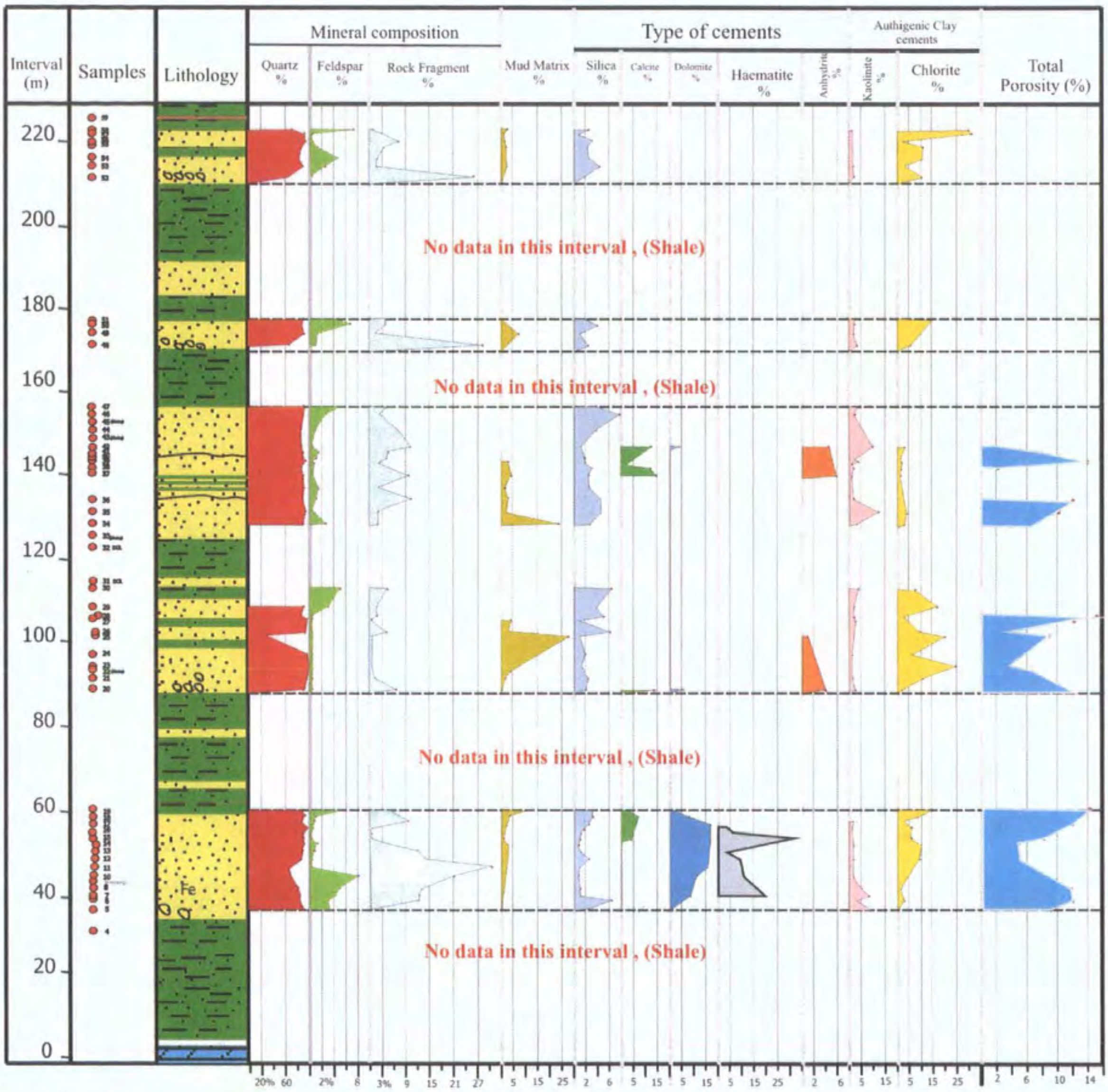


Fig. 4.2. Petrographic characteristic of Abu Shaybah Formation from Wadi Ghan section.

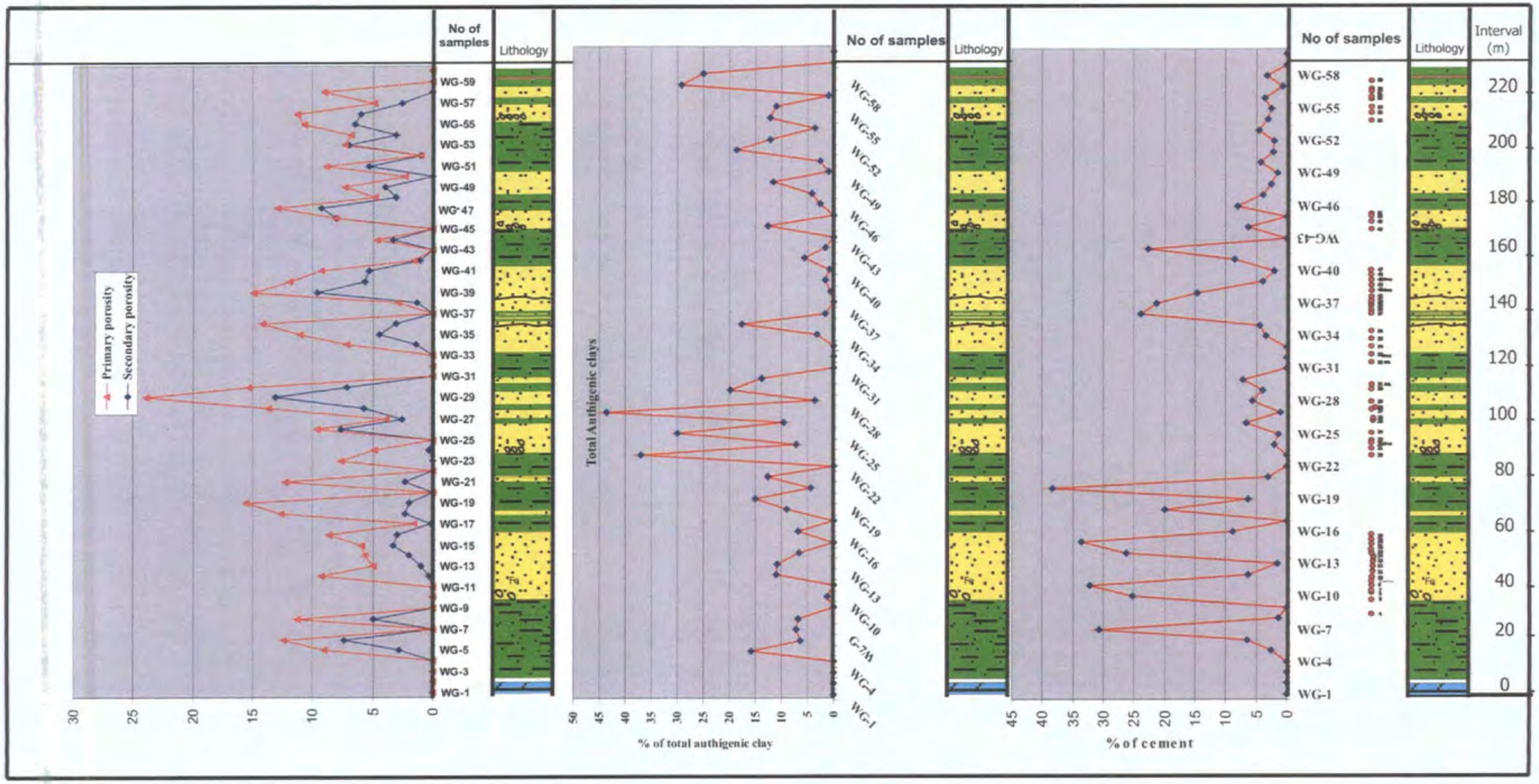


Fig. 4.3. Relationship between lithology and total cements, total authigenic clays and total porosity distributions in Abu Shaybah sandstones.

4.3.1 Petrography of Quartz arenite (Fig. 4.4)

14 thin sections have been analyzes used in the sandstone in this rock and the main detrital grains of the quartz arenite are tabulated below;

Detrital grains	**Percentage (%)**
Monocrystalline quartz	66.8
Polycrystalline quartz	1.2
Lithic fragments (mainly sedimentary rock fragments)	0.98
Micas (muscovite)	0.3
Heavy minerals, (mainly tourmaline)	0.44

Table 4.2. Main component of quartz arenite.

Type of total cements in these samples is:

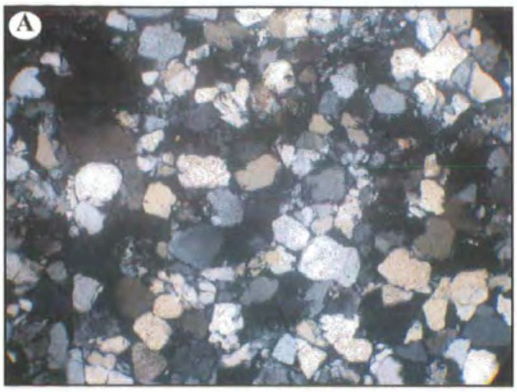
Detrital cements	**Percentage (%)**
Quartz overgrowth	2.6
Haematite	0.5
Anhydrite	0.45
Calcite	1.15

Table 4.3. Type of cements in quartz arenite.

Type of authigenic clay is:

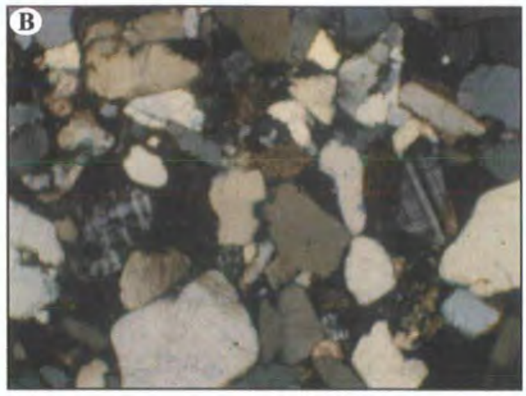
Detrital clays	**Percentage (%)**
Chlorite	11
Kaolinite	2.3
Mud matrix	1.3

Table 4.4. Percentage of authigenic clays and mud matrix in quartz arenite.



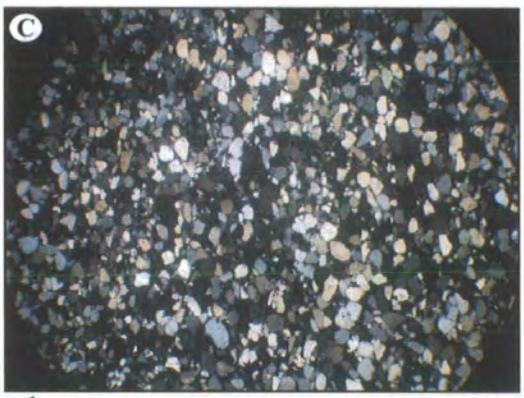
4mm

General view of quartz grains. WG-5, X4



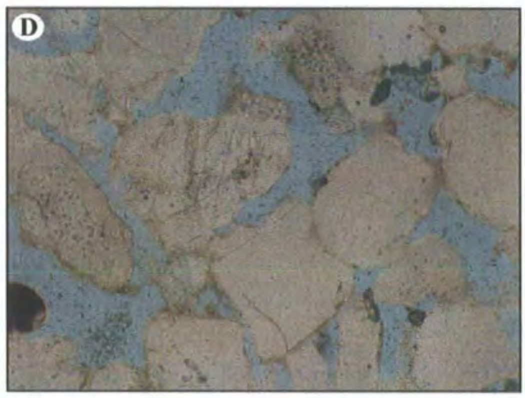
2mm

Feldspar grain, microcline. WG-5, X10



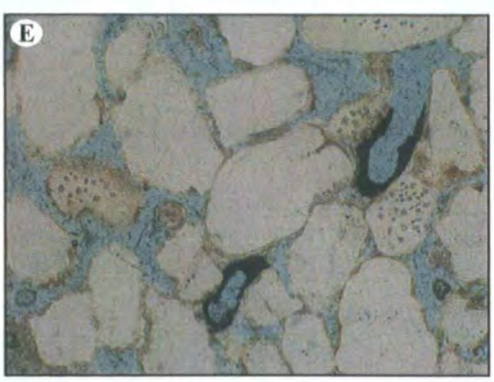
2mm

General view of quartz grains. WG-21, X2



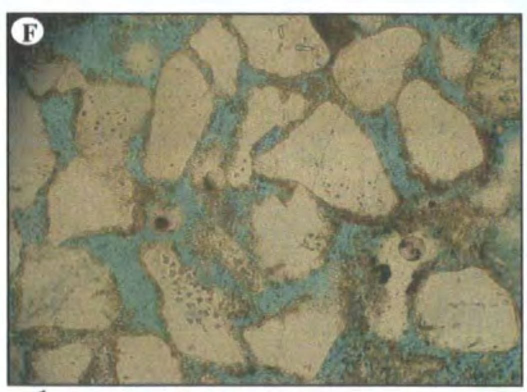
2mm

Good primary and secondary porosity of sub angular to subrounded quartz grains. WG-21, X2



2mm

Grain dissolution porosity. WG-21, X10



2mm

Authigenic chlorite pore lining in sub angular monocrystalline quartz grains. WG-29, X10

Fig.4.4. Showing textures, composition and type of porosity of quartzarenite.

4.3.2 Petrography of Sublitharenite (Figs. 4.5, 4.6 and 4.7)

24 thin sections had been classified in this type and the main composition is below:

Detrital grains	**Percentage (%)**
Monocrystalline quartz	61.8
Polycrystalline quartz	4.19
Lithic fragments (mainly sedimentary rock fragments)	2.17
Feldspar grains, mainly plagioclase	1.41
Micas (muscovite)	0.5
Heavy minerals, (mainly tourmaline)	0.69

Table 4.5. Main component of Sublitharenite.

Type of total cements in these samples is:

Detrital cements	**Percentage (%)**
Quartz overgrowth	3.44
Hematite	2.5
Anhydrite	0.38
Calcite	2.16

Table 4.6. Type of main types of detrital cements in Sublitharenite

Type of authigenic clay is:

Detrital clays	**Percentage (%)**
Chlorite	4.98
Kaolinite	2.32
Mud matrix	1.2

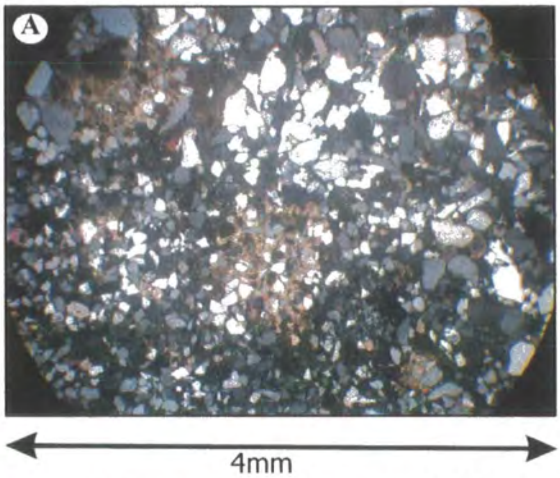
Table 4.7. Type of authigenic clays in Sublitharenite

Porosity characteristic

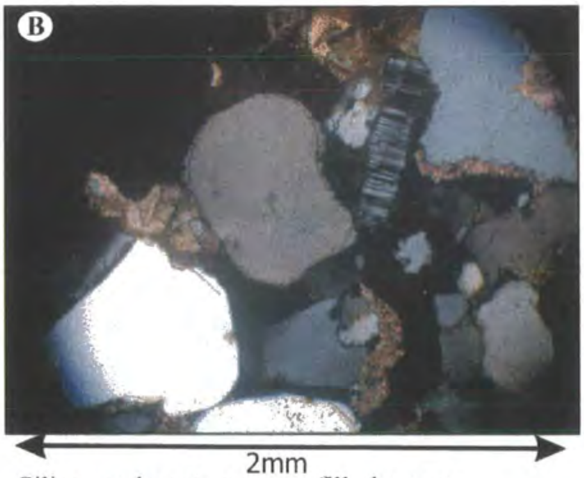
Two type of porosity identified in this type of samples:

Type of Porosity	**Percentage (%)**
Primary	3.5
Secondary, mainly grain dissolution	4.2

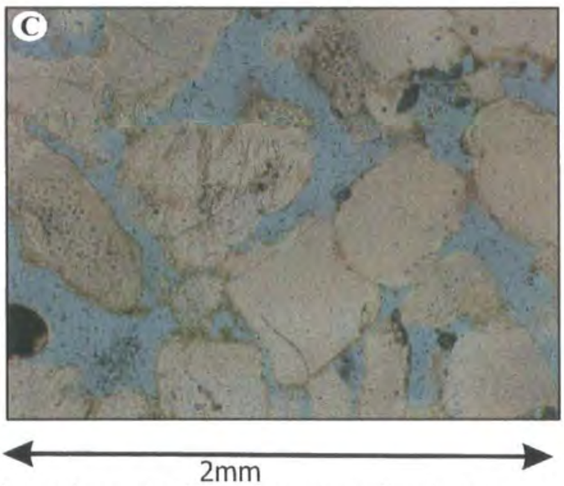
Table 4.8. Type of porosity in Sublitharenite



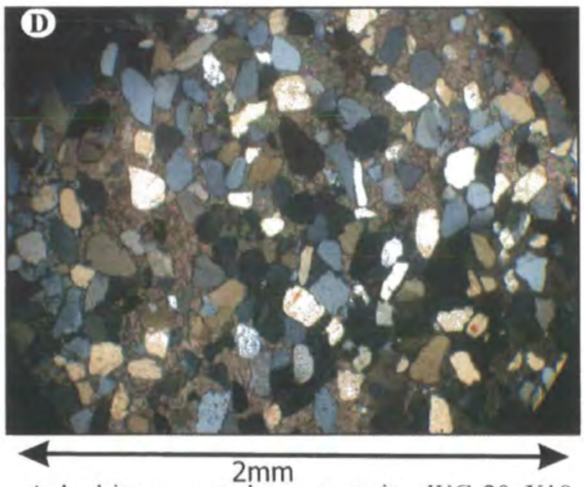
General view of quartz grains. WG-19, X2



Silica , carbonate cements fill the pores of quartz and plagioclase grains . WG-19, X10



Secondary porosity, due to grain dissolution of unstable minerals, shaly rock fragments. WG-19, X10



Anhydrite cemented quartz grains. WG-20, X10

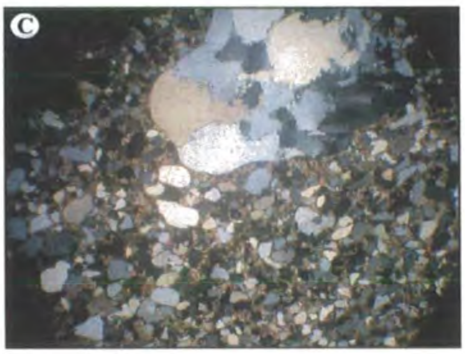
Fig.4.5. Textures, composition, porosity and type of cements in Sublitheranite.

4mm

Metamorphic rock fragments (MRF), in carbonate cement. WG-12,X2 ,XN

4mm

Same as previous photo but under PPL.



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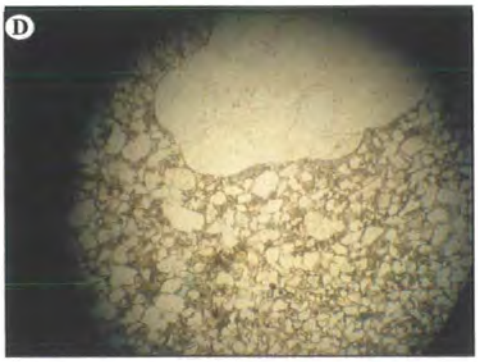
Polycrystalline quartz grains, in carbonate cement. WG-17,X2



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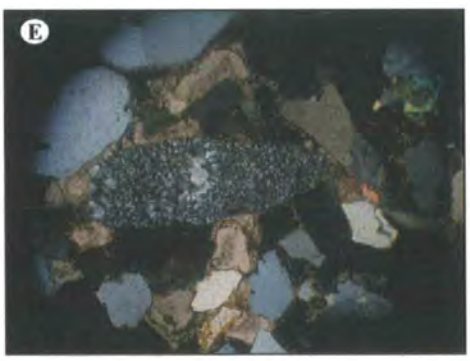

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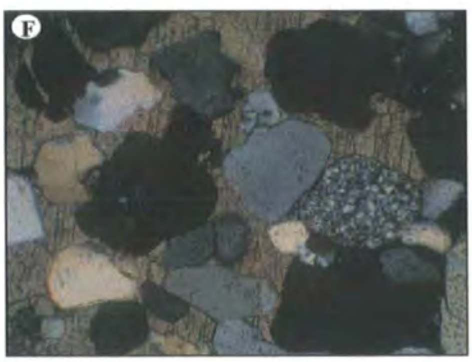
Sedimentary rock fragment floating in carbonate cement. WG-17, X10



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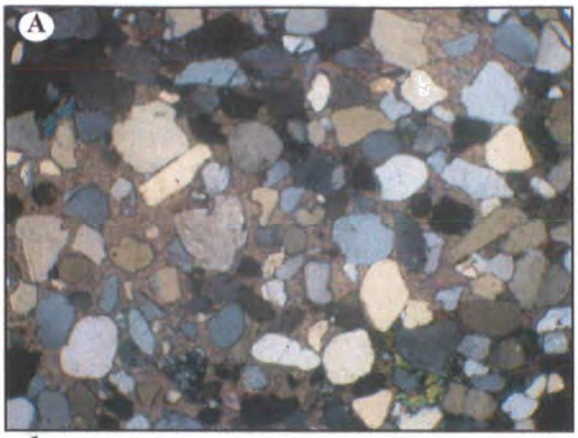
Sedimentary rock fragment and mono quartz grains floating in carbonate cement. WG-38, X10

Fig. 4.6. Type of rock fragments in Sublitheranite.



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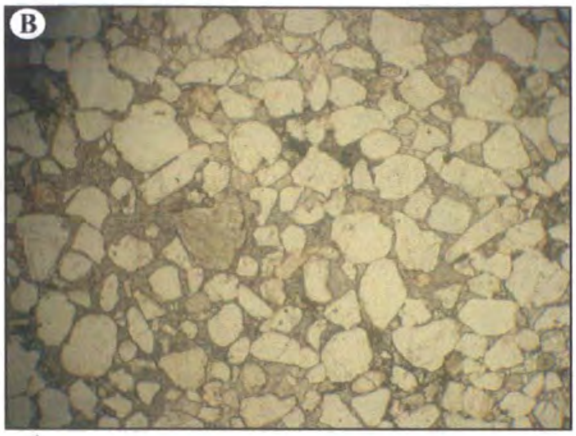
Sublitharenite cemented by carbonate .
WG-38, X10



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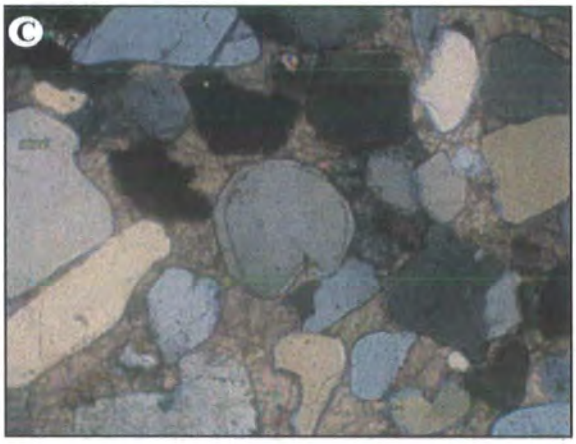

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Photo 3.3 Silica growth on the quartz grain surrounding by carbonate cemented.
WG-38, X10

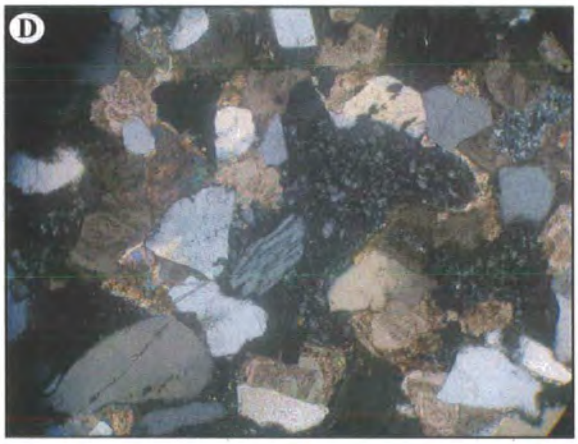

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Photo 3.4 Authigenic kaolinite are filled the pore space. WG-7, X10

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Photo 3.5 Quartz grains changed into carbonate cemented. WG-38; X10

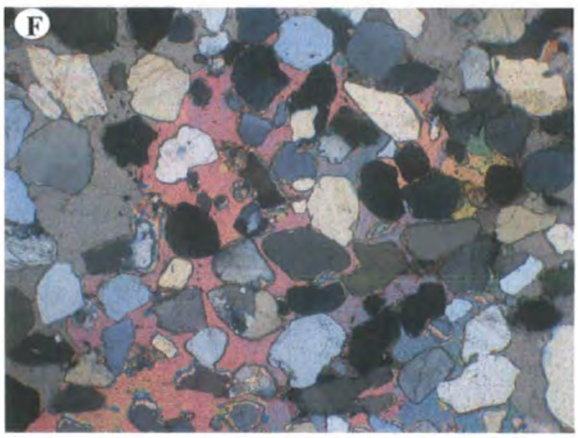

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Photo 3.6 Anhydrite cemented quartz grains.
WG-43, X10

Fig. 4.7. Type of rock fragments in Sublitheranite.

4.3.3 Petrography of Litharenite (Fig. 4.8)

Four thin sections have been classified in this type and the main composition is below:

Detrital grains	Percentage (%)
Monocrystalline quartz	52.07
Polycrystalline quartz	16.9
Lithic fragments (mainly sedimentary rock fragments)	4.13
Feldspar grains, mainly plagioclase	1.65
Micas (muscovite)	0.5

Table 4.9. Showing main detrital grains in Litharenite.

Type of total cements in these samples is:

Detrital cements	Percentage (%)
Quartz overgrowth	1.4
Anhydrite	5.7
Dolomite	8.5

Table 4.10. Type of total cements in Litharenite.

Type of authigenic clay is:

Detrital clays	Percentage (%)
Chlorite	4.67
Kaolinite	1.6
Mud matrix	1.4

Table 4.11. Type of authigenic clays in Litharenite.

Porosity characteristic

Two type of porosity are identified in this type of samples:

Type of Porosity	Percentage (%)
Primary	1
Secondary, mainly grain dissolution	0.8

Table 4.12. Type of porosity in Litharenite.

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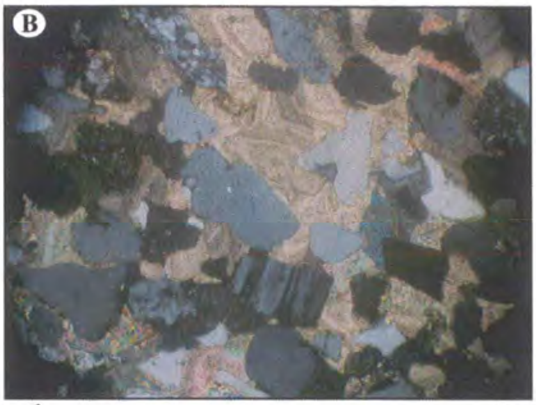
General view of mono quartz , polycrystalline quartz grains, feldspar plagioclase, carbonate cement. WG-10,X2 ,XN



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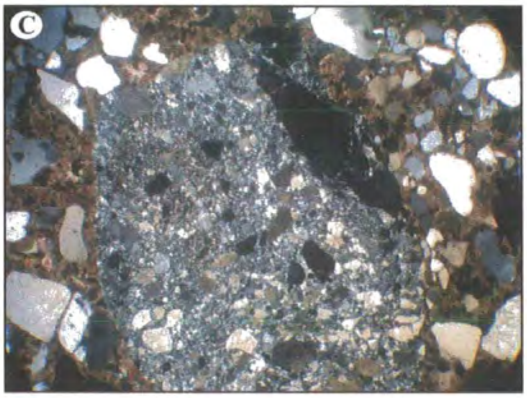
General view of mono quartz , polycrystalline quartz grains, feldspar plagioclase, carbonate cement. WG-10,X10 ,XN



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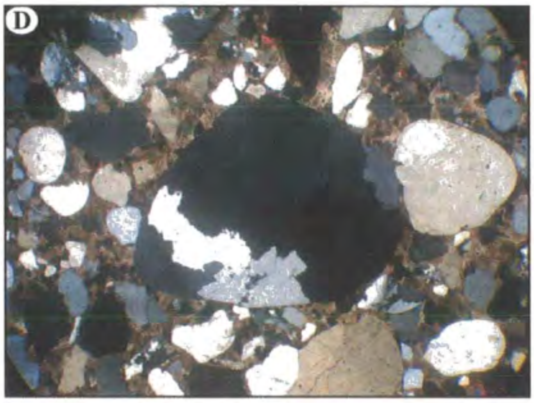
Sedimentary rock fragments (SRF), in carbonate cement. WG-11,X2 ,XN



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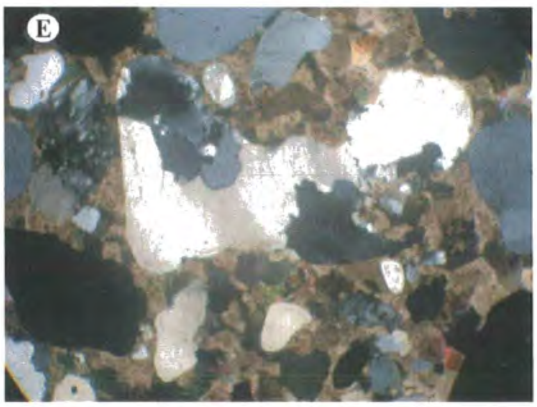
Polycrystalline quartz grains, in carbonate cement. WG-11,X4 ,XN



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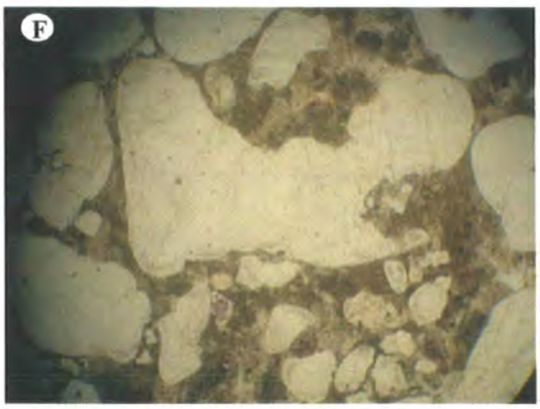
Polycrystalline quartz grains altered to carbonate cement. WG-11,X2 ,XN



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Same as previous photo but under PPL.

Fig. 4.8. Type of rock fragments in Litheranite.

4.3.4 Petrography of Subarkose

Three thin sections had been classified as subarkose and the main composition is below:

Detrital grains	Percentage (%)
Monocrystalline quartz	57.9
Polycrystalline quartz	1.7
Lithic fragments (mainly sedimentary rock fragments)	1.9
Feldspar grains, mainly plagioclase	4.23
Micas (muscovite)	0.5
Heavy minerals are mainly tourmaline	0.83

Table 4.13. Type of main components in Subarkose.

Type of total cements in these samples is:

Detrital cements	Percentage (%)
Quartz overgrowth	1.8
Anhydrite	7.6
Dolomite	11.3

Table 4.14. Type of total cements in subarkose.

Type of authigenic clay is:

Detrital clays	Percentage (%)
Chlorite	6.2
Kaolinite	2.13
Mud matrix	1.5

Table 4.15. Type of authigenic clays in subarkose.

Porosity characteristic

Two type of porosity are identified in this type of samples:

Type of Porosity	Percentage (%)
Primary	1.4
Secondary are mainly grain dissolution	0.56

Table 4.16. Type of porosity in subarkose.

4.3.5 Petrography of lithicwackes

Two thin sections had been classified as lithicwackes and are described below:

Detrital grains	Percentage (%)
Monocrystalline quartz	55.1
Polycrystalline quartz	0.2
Lithic fragments (mainly sedimentary rock fragments)	1.7
Feldspar grains, mainly plagioclase	1.7
Micas (muscovite)	2.8
Heavy minerals are mainly tourmaline	0.2

Table 4.17. Type of main components in lithicwackes

Type of total cements in these samples is:

Detrital cements	Percentage (%)
Quartz overgrowth	3.4

Table 4.18. Type of total cements in lithicwackes.

Type of authigenic clay is:

Detrital clays	Percentage (%)
Chlorite	1.1
Kaolinite	2.2
Mud matrix	24

Table 4.19. Type of authigenic clays in lithicwackes.

4.4. Provenance (Fig. 4.9)

Sandstone composition can be a useful indicator for composition, climate and tectonic setting of the source area (Swift *et., al.*, 1991; Macdonald, 1993). Quartz arenites may originate as first cycle deposits (Pettijohn *et al.*, 1978). They can also be a product of multiple recycling of quartz grains from sedimentary source rocks or may have formed *in situ* through extensive chemical weathering. Quartz arenites are typically deposited in stable cratonic environments (Suttner *et al.*, 1981; Tucker, 1991; Boggs, 1995).

Arkoses of the ASF contain plagioclase and microcline derived from feldspar-rich rocks (4.23%) (e.g. granites or gneisses); they can also be derived from a granitic source area or can also be authigenic in origin, controlled by the rate of erosion and climatic setting. Arkoses can accumulate through tectonic uplift that brings granitic basement rocks up into the zone of erosion.

Litharenites of the ASF contain mainly sedimentary rock fragments, mudstone fragments and flakes of micas, mainly muscovite, derived from low grade metamorphic equivalents.

Litharenite may change through time reflecting uplift in the source area, and the availability of different rock types (Tucker, 1991).

Ternary diagrams (Fig. 4.9) have been used to plot the ASF composition data for the Wadi Ghan section (Fig. 3.1). These indicate that they were derived from Tibesti massif south of Libya which is composed of Precambrian rocks (Fig. 2.1). These rocks are strongly folded and unconformably overlain by a thick sequence of Palaeozoic clastic rocks (Gargaf Group). The ultimate source area of the ASF was thus located in the vast, uplifted continental area which in the Triassic extended from the Gargaf and Tibesti uplifts as far as the great Sirt uplift (Geological map of Libya, Fig. 2.1), and similar parent rocks (source rock). Moreover, the ASF in the study area, formed under arid or semi-arid climatic setting that become increasingly humid (Suttner *et al.*, 1981).

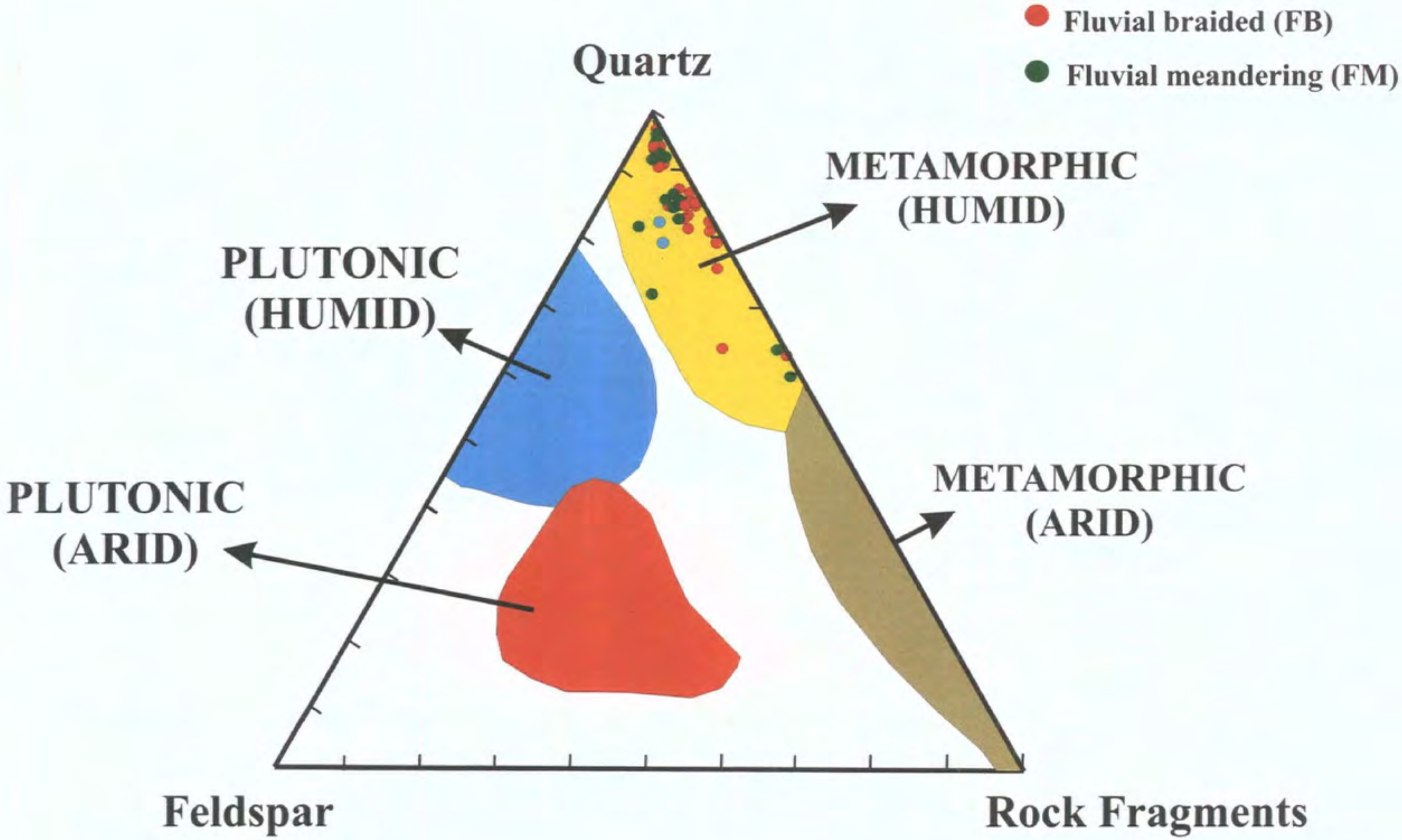


Fig. 4.9. Ternary QFRF Plot showing the average composition of sandstones and source rock marked by metamorphic humid climate from Wadi Ghan section (After Suttner *et al.*, 1981).

4.5. Diagenesis

These are many factors affecting the diagenesis of the ASF that have been observed from thin sections (Fig. 4.10). The depositional environment, climate, composition and texture of the sediment are initial controls and then pore-fluid migrations, the burial history and other factors affect the course of diagenesis.

4.5.1 Silica cementation (Figs. 4.4D, 4.5B, C and 4.7C)

Silica cement is an early authigenic cement and common diagenetic phase in the Abu Shaybah sandstones. Authigenic silica cement is present in the form of syntaxial overgrowths on detrital grains with well developed crystal faces, where clay rims are visible around detrital quartz grains. In many cases the shape of the original grain is delineated by a thin iron oxide or clay coating between the overgrowth and the grain (a dust line). A thicker clay rim around the quartz grains, precipitation of a syntaxial overgrowth.

The early diagenesis of silica has reduced the primary porosity. The origin of the silica for this cementation frequently has been attributed to pressure dissolution (Tucker 2001).

Quartz cement is a main factor influencing reservoir quality in siliciclastic sandstones. Quartzose sandstones often have quartz as the main cement mineral that destroys porosity and permeability (Molenaar, *et al.*, 2007).

4.5.2 Carbonate cementation, (Figs. 4.5B, 4.6A, B, C, F, 4.7A, C, D, E, 4.8A, B, C, D, E and F)

Calcite and dolomite are abundant in the lower part of the Abu Shaybah Sandstone as pore filling and replacement cements of post-depositional origin. The primary and early diagenetic carbonate cement of many modern types of sediment may include magnesium in calcite.

4.5.2.1. Calcite cements

Drusy calcite mosaics and poikilotopic calcite – crystals are common in grain-supported sandstones and filled the pore space between grains, such as quartzarenite, sublitharenite and arkoses.

The early precipitation of calcite inhibits later quartz overgrowth formation and feldspar alteration to clays but it can result in total loss of porosity and permeability

(Tucker, 2001). Precipitation of $CaCO_3$, taking place when the solubility product is exceeded, often occurs through an increase in the activity of the carbonate ion.

4.5.2.2. Dolomite cements

Dolomite cement in Abu Shaybah sandstone is common as well formed millimetre sized rhombohedra, and they are commonly iron rich (ferroan), indicating precipitation in reducing conditions. Magnesium for later dolomite precipitation may be derived from clays or dissolution of magnesium-rich silicates.

4.5.3 Clay mineral and cement, (Figs. 4.4 D, E and F).

Kaolinite and chlorite are the most common authigenic clay minerals developed within the Abu Shaybah Sandstones.

4.5.3.1. Kaolinite

Kaolinite is a late – stage authigenic mineral post-dating early quartz overgrowth. Kaolinite occurs as dispersed pore- filling cement between quartz grains or replacement of pre- existing detrital grains.

Kaolinite formed as vermicular booklet cements within the Abu Shaybah Sandstones, and may be precipitated from the pore-fluid or alteration of detrital feldspars (Franks and Foster, 1984).

Kaolinite is the characteristic mineral of acid fresh water environments (Pettijohn, 1957). Bjørlykke (1983) pointed out that fluvial sandstones and some shallow-marine sandstone may be flushed by meteoric water after deposition to form authigenic kaolinite.

4.5.3.2. Chlorite

Chlorite and smectite are usually associated as pore lining and pore filling phases. These minerals come from alteration of feldspars, and non ferroan calcite is common as a burial cement in these rocks too.

4.6. Diagenetic phases and implications

The spatial and temporal distribution of diagenetic alterations in siliciclastic sequences is controlled by a complex array of interrelated parameters that prevail during diagenesis. Morad *et al.* (2000) and Ketzer *et al.* (2003) widely discussed the significance of diagenetic phases for sequence stratigraphy. This study has further identified the significance of diagenetic phases for incised valley fills and understanding their evolution., but similar occurrence can be recognised at LST, HST and SB. This is of particular importance when types to correlate fluvial succession and has major implication for hydrocarbon exploration and the understanding of how these fluvial successions develops through tectonic and are affected by climate change. It has been identified in this study that the spatial and temporal variations in diagenesis are controlled by a complex array of interrelated reactions in siliciclastic successions. The spatial distribution as identified along section of the ASF is controlled by climate, detrital composition, under lying and overlying marine sediments (Al Abu Ghaylan Aziziyah Formations) relative sea-level changes and ground water penetration. However, the most important alteration in fluvial sediments, such as the ASF, is silicate dissolution and the formation of kaolinite (Figs. 10 and 11).

4.6.1. Diagenetic alterations associated with LST and sequence boundary

The dissolution of framework grains and formation of kaolinite in the LST of the lower sequence has possibly occurred during near-surface diagenesis, i.e. before significant compaction, as kaolinitized micas and the collapsed texture of extensive kaolinitized feldspar grains. These sandstones are more prone to kaolinitization and grain dissolution compared with TST and HST estuarine, because of more effective exposure to meteoric water circulation during near –surface diagenesis (Morad et al., 2000). The humid climate conditions that prevailed during deposition enhanced meteoric water percolation and dissolution of feldspars, micas and lithic fragments, and the formation of kaolinite (Ketzer *et al.*, 2003).

In addition to the formation of kaolinite in the LST deposits of the lower sequence, local increase in kaolinite content in the lag deposit of the upper sequence and below the upper sequence boundary is attributed to possible subaerial exposure during fall in relative sea level (regression) and meteoric water flushing of marine sediments (Morad *et al.*, 2000; Ketzer *et al.*, 2003).

4.6.2. Diagenetic alterations associated with TST, HST and MFS (Fig. 4.11)

The formation of calcite and kaolinite, and grain dissolution, occurred mainly in sandstones of the TST and HST in the vicinity of parasequence boundaries, transgressive surface and maximum flooding surface.

Grain dissolution and kaolinite formation below parasequence boundaries (Figs. 3.11 and 4.10), by meteoric water changed with CO2 and organic acids as a result of percolation through peat deposits (Staub and Cohen, 1978). Chlorite formation in such sediments is usually attributed to mesogenetic transformation of iron-rich, clay precursors such as berthirine and odinite, which are common eogenetic minerals in paralic and shallow-marine settings (Odin and Matter, 1981).

Calcite

Calcite in the ASF occurs mainly as strata bound cement along parasequence boundaries. Such a distribution pattern of calcite cementation at parasequence boundaries is usually attributed to the increased amount of carbonate bioclastic (South and Talbot, 2000; Ketzer *et al.*, 2002).

The relationship between calcite and other diagenetic minerals indicates that cementation commenced at near sea floor conditions, with microcrystalline texture, and continued as mosaic and poikilotopic crystals during these phases of cementation (Ketzer *et al.*, 2003).

Chlorite

Chlorite is a minor and rare consistituent of the studied sandstones (Table 4.2). It occurs as grain coating platelets arranged perpendicul to the grain surfaces, as pore filling aggregates showing a rosette-like texture or as scattered platelets within pseudomatrix and micas. Chlorite was detected in whole samples but concentrated in HST and near MFS).

Kaolinite

Kaolinite occurs as booklets and vermicular aggregates (Fig. 4.2) that fill intergranular pores or replace micas. Kaolinite is abundant in incised valley fill in the LST.

Quartz

Quartz cement occurs as syntaxial overgrowths around detrital quartz grains. Diagenetic quartz occurs as a minor constituent in sandstones of all systems tracts.

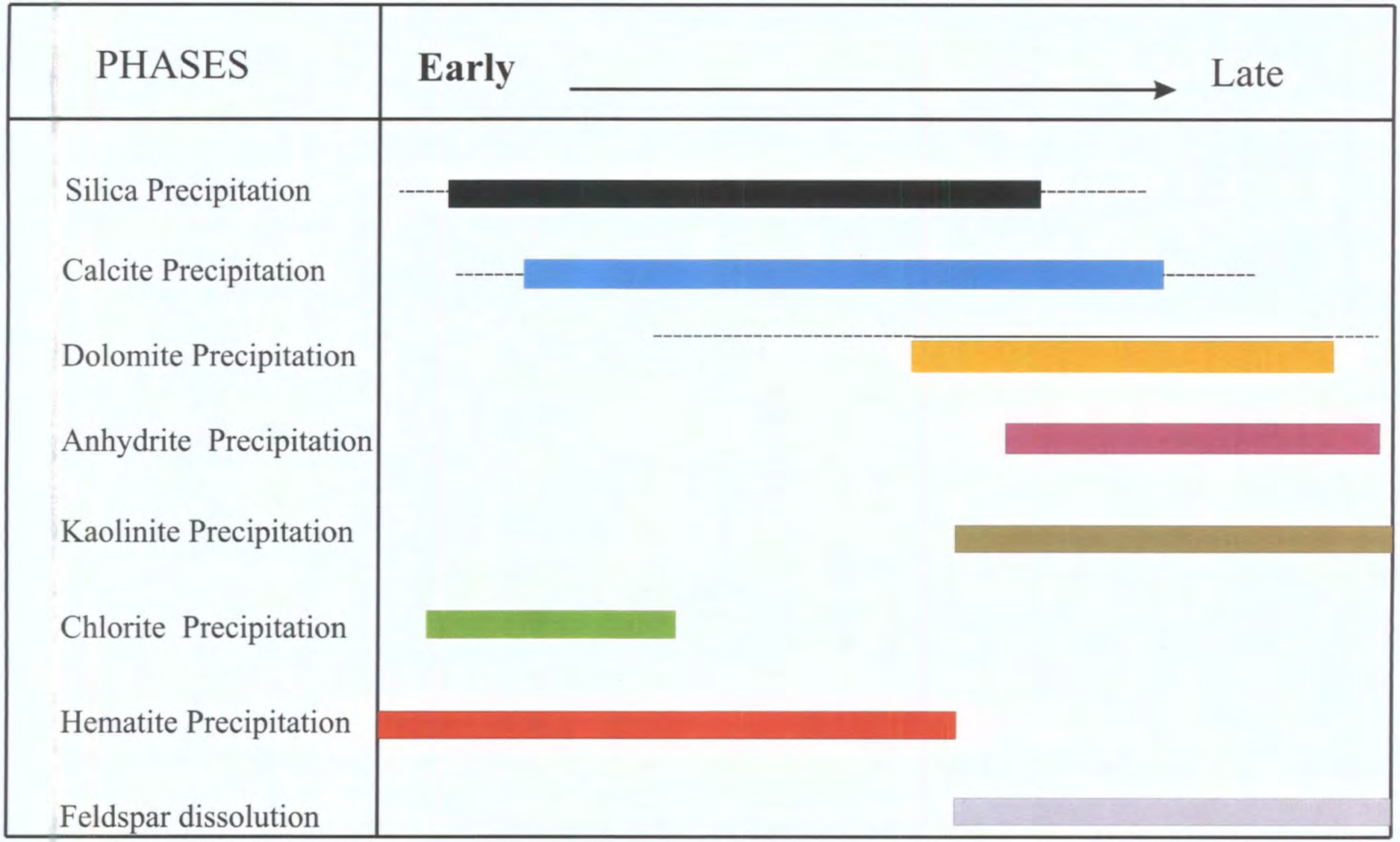


Fig. 4.10. Showing Diagenetic phases of Abu Shaybah Formation (ASF), Wadi Ghan section.

Feldspars

Diagenetic feldspar is a minor constituent in sandstones of all systems tracts and occurs as thin (10-20 μm) K-feldspar grains and small (40 μm) prismatic crystals replacing detrital plagioclase grains. Diagenetic feldspars show no obvious distribution pattern in relation to facies or sequence stratigraphy (Ketzer *et al.*, 2003).

Dolomite

The formation of microcrystalline dolomite was followed by the precipitation of zoned dolomite crystals in open pores as well as in pores formed by the dissolution of calcite bioclasts. The ferroan composition of microcrystalline dolomite and zoned dolomite indicates that dolomitization occurred by modified seawater or brackish water, in a reducing, nonsulfidic environment (Berner, 1981).

Other diagenetic minerals

Other diagenetic minerals occur as traces and include haematite concentrated near the sequence boundaries in the Low stand system tract (LST) in incised valley fill, and anhydrite in the HST sequence.

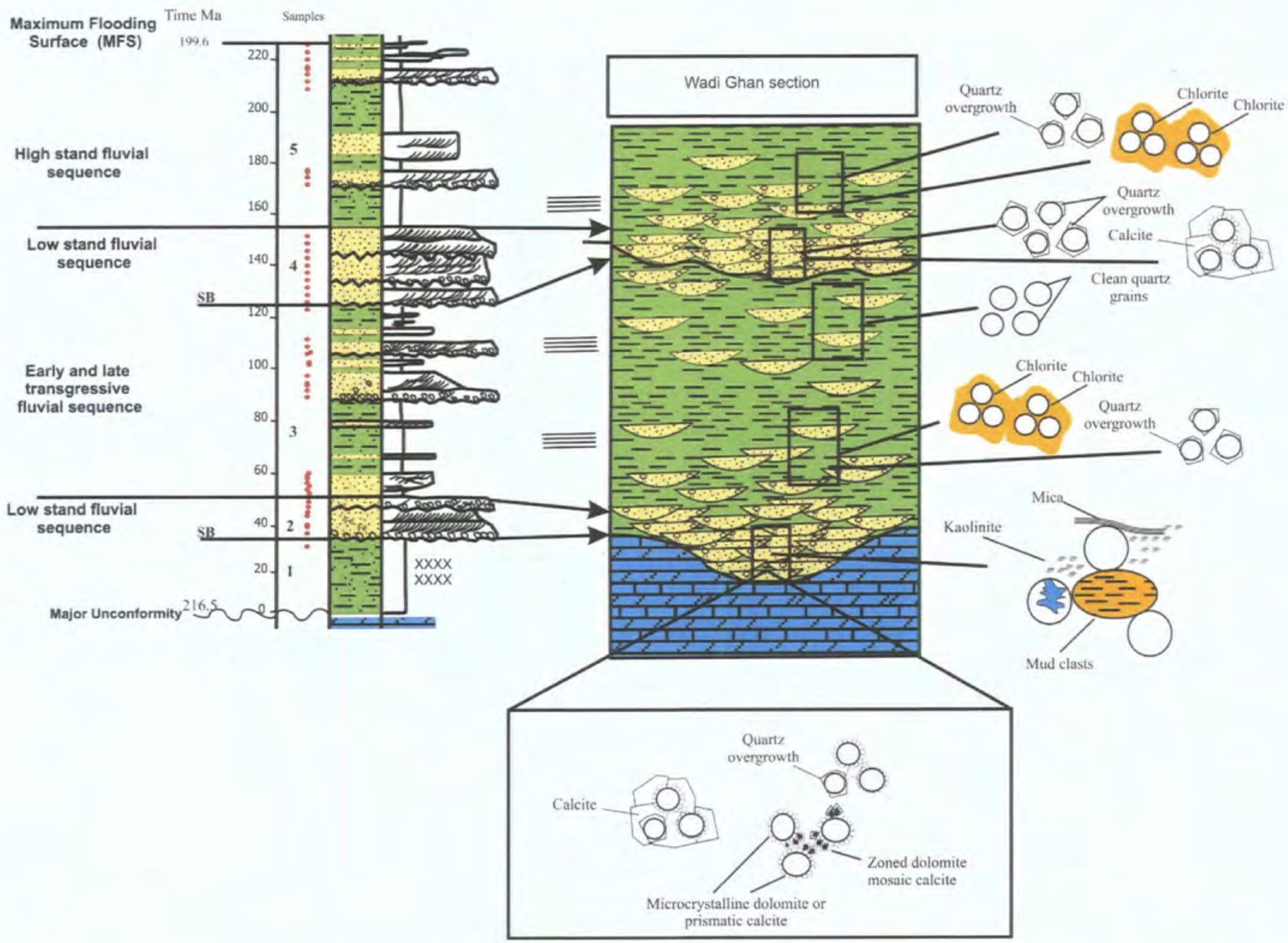


Fig. 4.11. Distribution of diagenetic alterations in low stand fluvial sequence, early and late transgressive fluvial sequence and high sand fluvial sequence of ASF.

4.7. Comparison and equivalent Triassic rocks in other Libyan Basins (e.g. Sirt, Murzuq, Al Kufrah and Ghadames Basins)

Triassic rocks have a wide distribution in different sedimentary basins in Libya (Fig. 1.7) and evidence of accumulation during the rifting phase in the Sirt Basin began in the Triassic, and the Triassic sediments are also found up to 200 m thick in shallow grabens in the Hamiumat Trough in the Northeast Sirt Basin (Abadi 2002). Surfaces and subsurfaces Triassic rocks in the Murzuq basin Zarzaitine Formation in about 400 m thickness, consists of multi-coloured claystone and fine grained sandstone in the lower member , mottled claystone , argillaceous limestone and dolomites with thin bands of gypsum and anhydrite and the upper part unit of friable, argillaceous sandstones claystone, dolomites and gravels (Hallett, 2002).

The Late Triassic rocks (ASF) in the Al Kufrah Basin are up to 50 m thick and consist of red sandstone and claystone. The sandstone consists trough cross bedding and floodplain deposits and up to 500 m in some wells are found in the basin, comparing continental siltstones and sandstones, with claystone.

On the southern flank of the Ghadames basin, up to 150 m of alternating bands of gypsiferous and ferruginous claystone, dolomitic marls, micaceous siltstones, quartzose sandstones and gravels, with a distinctive brown colour.

In the offshore Sabratah Basin, Triassic rocks consists of a thick series of halites Piercement salt-domes are present in the area of Ra's Ajdir (Libyan –Tunisia border).

4.8 Discussion

4.8.1. Provenance

The ASF had its source in siliciclastic sedimentary rocks, as indicated by their high compositional maturity, the predominance of monocrystalline quartz, which commonly bears inherited overgrowths, feldspar and mica, and the scarcity of sandstone rock fragments.

Abu Shaybah Formation has an average framework composition of Q88.3 F9.8 R1.9 and 79.7% of the quartz grains are monocrystalline. The lower ASF are composed of quartz-arenites and sublitharenite were derived from Palaeozoic siliciclastic rocks, whereas middle and upper part of ASF there is an increase in polycrystalline quartz grains and it is mainly sublitheranite, quartzarenite and rare subarkose originated mainly from metamorphic terrains. This change in provenance has contributed to the significantly different diagenetic paths followed by the Lower ASF quartz arenite and sublitheranite and middle and upper ASF sublitheranite, quartz-arenite and rare subarkose. Grain coating chlorite clays are abundant in the upper ASF, where they exert a critical control on reservoir quality. Whereas the lower ASF has abundant carbonate cement (dolomitic), plus abundant hematite in the lower ASF which is absent in the middle – upper ASF.

4.8.2. Use of Sequence stratigraphy to understanding the distribution of diagenesis in fluvial deposits

Diagenesis and sequence stratigraphy have been formally treated as two independent disciplines in sedimentary geology. It was just recently documented that diagenetic alterations such as calcite and dolomite cementation might be distributed at sites coincident with sequence stratigraphic surfaces (Tucker, 1996; Taylor and Curtis, 1995).

Sequence stratigraphy, if applied independently, allows facies prediction, and thus provides information on depositional-related reservoir quality (Van Wagoner *et al.*, 1990; Posamentier and Allen, 1999), which is controlled by sorting and grain size.

Shallow diagenetic alterations in the ASF, which include formation of kaolinite, infiltrated clays, iron oxide, and carbonates are encountered in all depositional facies and system tracts. Nevertheless, the amounts and the spatial and temporal distribution of these alterations vary fairly systematically among the depositional facies and system tracts, reflecting primarily variations in parameters related to climatic conditions and changes in pore water chemistry in response to changes in the relative sea level. Conversely, mesogenetic ankerite, dolomite, albitized feldspars, and chlorite do not display systematic variation among the depositional facies and system tracts, whereas the distribution of mesogenetic silicates, (i.e. illite, kaolinite, and quartz overgrowths) within the sequence stratigraphy is apparently controlled by the distribution of diagenetic alteration (El-ghali *et al.*, 2006).

Sandstones of ASF in Jabal Nafusah (sequence boundaries I and II) display a variety of diagenetic features in thin section, which reflect the effects of mechanical compaction, dissolution of framework grains, and precipitation of clay filtrated and coating the detrital grains, authigenic minerals chlorite, kaolinite, quartz overgrowth, and carbonate), (Fig. 4.10). These diagenetic processes are responsible for destroying much of the primary intergranular porosity and permeability in the ASF.

Application of sequence stratigraphy to the ASF allows recognition of two sequence boundaries I to II (Fig. 4.11). Detailed comparison of the relative abundance of major detrital framework grains within each depositional sequence provides a explanation of the compositional variation in the ASF. Sandstones above sequence boundary I typically have a higher abundance of monocrystalline quartz, feldspar and rock fragment and cemented by calcite, dolomite, and clay filtrated and coating the quartz grains, the composition of framework grains varies systematically from the lowstand system tract to highstand fluvial sequences within each depositional sequence. Lowstand system tract sandstones contain more chemically unstable detrital framework grains such as abundant kaolinite and grain dissolution owing to a more effective meteoric water circulation and contemporary humid climatic conditions.

Grain-coating clays

Grain –coating clays, which occur as platelets that are tangentially arranged around the detrital grains surfaces and, which rarely have a grain bridging texture, suggest formation by mechanical infiltration (Matlack *et al.,* 1989; Moraes and De Ros, 1990, 1992). The occurrence of grain-coating rather than-bridging clays indicates that infiltration occurred within the phreatic zone, close to a fluctuating water table (Moraes and De Ros, 1990; Ketzer *et al.,* 2003). Infiltrated clays in fluvial incised-valley LST sandstone were likely formed during a late stage of on LST, when accommodation is progressively created in the fluvial environments (Shanley and McCabe, 1994). Accommodation creation in fluvial incised-valley setting was enhanced by relative sea level rise due to fluvial retreat, glacial (El-Ghali, 2005) and thus has resulted in increasing deposition of floodplain mud (Wright and Marriot, 1993), repeated flooding and infiltration of mud-rich waters into subaerially exposed, permeable sand.

Kaolinite

The dissolution of unstable framework grains such as micas, mud intraclast and feldspars as well as mud matrix and pseudomatrix, and the formation of kaolinite in the ASF is attributed to near surfaces, meteoric water diagenesis (Morad *et al.,* 2000; Ketzer et al., 2003a, b) and /or to flux of organic acids during diagenetic origin of kaolinite includes : (i) expanded texture of kaolinitized micas, which typically implies their occurrence in poorly compacted sandstones prior to significant burial (Ketzer *et al.,* 2003a); (ii) vermicular texture of kaolinite, which has been suggested to occur soon after deposition (McAulay et al., 1994; Osborne et al., 1994; Wilkinson et al., 2004); (iii) close association of kaolinite with diagenetic, Mg-poor siderite; and (iv) partial transformation of kaolinite into dickite and illite. Kaolinite in the incised – valley LST sandstones (trace – 9.5 %, av. 2 %).

Quartz overgrowths

Quartz cement, which displays syntaxial overgrowths, and which is associated with sites of intergranular dissolution, is typically of mesogenetic origin (Morad *at el.*, 2000). The distribution of quartz overgrowths within depositional facies and systems tracts is strongly controlled by the spatial and temporal distribution of infiltrated coating clays and early calcite and siderite cements.

Calcite cements

Calcite cement, which occurs in all depositional facies and system tracts (traces – 7.3%, av. 0.8% in Lowstand fluvial sequence boundary I, (traces – 15%, av.3.3% in Lowstand fluvial sequence boundary II) , which occurs as blocky aggregates, replaces and displaces the framework grains and tends to fill large pores between loosely packed framework grains.

Dolomite cements

Dolomite cement, which occurs on a lower sequence boundary in a low stand fluvial sequence (trace – 22.3% av. 6.1%), exhibits rhombic textural habits.

4.9. Conclusion

Specific conclusions from this chapter include:

- There is a significant change in detrital composition between the lower – middle and upper ASF, caused by a change in provenance. The lower ASF quartz arenites and sublitheranite were derived from Palaeozoic siliciclastic rocks, whereas middle and upper part of ASF quartz grains show an increase in polycrystalline quartz grains and are mainly sublitheranite, quartzarenite and rare subarkose originated mainly from metamorphic terrains. This change in provenance has contributed to the significantly different diagenetic paths followed by the Lower ASF quartz arenite and sublitheranite and the middle and upper ASF sublitheranite, quartzarenite and rare subarkose.
- Grain coating chlorite and locally kaolinite clays, common in the ASF, originated from clay infiltration, and facies-related variations in the volume of grain-coating clay.
- The precipitation of authigenic vermicular kaolin it was related to the dissolution of detrital feldspar.
- Quartz cement is one of the most common cements in whole interval of ASF.
- Application of sequence stratigraphy to the late Triassic of Abu Shaybah Formation allows recognition of two sequence boundaries. Detailed comparison of the relative abundance of major detrital framework grains within each depositional sequence provides a more sophisticated explanation of the compositional variation in the ASF. The formation of grain-coating clays and kaolinite, which has occurred by percolation of mud-rich surfaces waters into sandstones of fluvial incised – valley LST sandstones during late stage of lowstand.
- The authigenic kaolinite distribution throughout the ASF section is attributed to the occurrence of a warm and rather wet climate.
- The detrital and authigenic mineral composition of ASF indicate that significant climatic fluctuations between relatively wetter and drier conditions appear to have occurred during the accumulation of the ASF Formation. Wetter climatic conditions are interpreted for the arenites with higher amounts of mono- and/or polycrystalline quartz, kaolinite, and quartz overgrowth. Drier climatic conditions are inferred for the arenites with lesser percentages of quartz but higher amounts of authigenic kaolinite.

Chapter 5

Application of Fluvial Architecture: Understanding Low–Net to Gross Reservoirs

Chapter 5

Application of Fluvial Architecture: Understanding Low–Net to Gross Reservoirs

5.1 Introduction

Nonmarine sequence stratigraphic models have largely been developed from studies of well-exposed foreland-basin successions in Utah and Wyoming (Shanley & McCabe, 1994; Van Wagoner,1995). These models suggest major unconformities, or sequence boundaries, are associated with a dramatic change in fluvial style from isolated single story, meandering streams in the transgressive and highstand systems tract to amalgamated braided-stream systems in the lowstand systems tract. However, there has been a significant challenge to this bipartite classification of ancient fluvial systems into 'meandering' versus 'braided' end members, in large part because these are not mutually exclusive categories (Bridge 1993, 2003). In modern systems, water discharge, gradient, sediment flux and grain size determine channel patterns. Gradational differences within and among these variables result in a continuum of channel patterns (e.g. Frostick & Jones 2002; Bridge 2003). Furthermore the interaction of climate and tectonism has long been recognised as an important control on fluvial sedimentation, but in recent years the recognition that steady-state processes are more fundamental for shaping fluvial systems than catastrophic has led to a reappraisal of fluvial sediments in line with climatic controls (Jones, 2004; Meadows, 2006). The understanding of these controls has important implications when modelling fluvial reservoirs for correlation purposes and imputing the correct parameters especially in low net to gross reservoirs.

The integration of diagenesis and sequence stratigraphy allows a better prediction of the spatial and temporal distribution of diagenetic alterations in siliciclastic sequences, and containing of reservoir quality evolution of sandstone. Diagenetic minerals such as calcite, dolomite, kaolinite and mechanical clay infiltration, and intergranular porosity showed a systematic distribution in sandstones lying in the vicinity of sequence and parasequence boundaries, transgressive and maximum flooding surfaces, and in sandstones of the lowstand, transgressive, and highstand systems tracts (Chapter 4).

Considering the above, this chapter explores the external geometry of channel sand bodies in low net to gross fluvial systems such as the ASF, such as approach requires on understanding of channel bodies them sections as reported from chapter 3. The will be used to understand the ASF as a reservoir analogue and compared unit the Triassic TAG-I fluvial sandstones of Algeria (Section 5.6.1). The chapter will finally discuss the implication for modelling subsurface fluvial data sets. Using correlation and sequence stratigraphic ASF incised valley system as a rest bed (section 5.5).

5.2. Qualitative terms to describe channel deposits; (Channel bodies and valley fills in ASF)

Fluvial channel deposits comprise a suite of widely recognized components. A set of bedforms, such as dunes and ripples, is typically organized into bars and bedload sheets, which lie within channels (Bridge 1993; Lunt *et al.*, 2004). Channel banks commonly migrate laterally as the adjacent floodplain deposits are eroded, with concomitant lateral migration of bars within the channel, and the channel base may incise into the underlying floodplain deposits, resulting in stacked bar deposits and bedload sheets. In this study, the channel generates a channel body (Fig. 5.1). This is larger than the original channel dimensions and is common occurrence in the ASF of the study area (Table 5.1).

Individual channel bodies in the ASF can amalgamate to form a composite channel body through the relocation (avulsion) of the river channel and juxtaposes against younger and older channel bodies (Fig. 5.1) Pettijohn et al., (1972) used the term belt to describe coalesced smaller bodies, but it is now widely used as a synonym for a channel body, that is used in this study.

Channel bodies can be differentiated into simple and complex types, which are readily identifiable in the Wadi Ghan section, Jabal Nafusah. Complex units are those devisable into more than one component on the basis of erosional discontinuities and scoured bases. These complex channels typically have multi-storey units with irregular basal scours in contrast to the simple, single-story. The terminology that is widely ascribed and used in this study is highlighted in Fig. 5.1.

The time required to generate a channel body may range from decades in the case of the most frequently avulsing systems (Sinh *et al.,* 2005) to tens of millions of years in the case of thick palaeovalley fill sequences with prolonged sediment transfer from active mountain belts (Jones *et al.*, 2001; Jones 2004). This presents the need of knowing the wider context of the fluvial system and the basinal context.

The reservoir geologist is particularly interested in how discrete channel bodies amalgamate either partly or fully to form a composite body with varied degrees of connectedness and often their description becomes a matter of judgement and experience. Settings that are likely to promote higher degrees of connectedness and concentration include sites where drainage patterns are forced to converge around growth structures (e.g. Jones, 2004), restricted points of entrance to a depositional basin (e.g. Hovius 1996; Jones *et al.*, 2001) and basins with differential subsidence (e.g. Gawthorpe and Leeder, 2000). The degree of interconnectedness of channels has been widely incorporated in computer models, but low net to gross fluvial systems have largely been neglected.

For the purpose of this study the outcrops of the ASF were grouped according width to thickness ratios (W/T). (i) Channel sandstone bodies of fluvial braided (Ribbon dominated), (ii) sandstone bodies of fluvial meandering (single-story-channel-fill, multistory-channel-fill) and (iii) Crevasse splay sand bodies.

5.2.1. Channel sandstone bodies of fluvial braided

Eleven channel sandstone bodies (Figs. 5.2 and 5.3b) characterized by gravelly sandstone bodies, poorly sorted, and poorly channellized systems with mainly planar and trough cross-stratification (GSp & GSt). These are overlain by climbing ripples, abundant mud clasts at the base, erosional surfaces and a range of widths from 50-200 m, (av. 170m) and thickness from 1.5 -5 m (av. 3 m). Most of these examples belong to the categories of gravel-dominated rivers and sand-dominated low sinuosity rivers recognized by Miall (1996).

5.2.2. Sandstone bodies of fluvial meandering (sheet dominated)

Twenty eight sheets of channel sandstone bodies with a sandy or heterolithic nature are characterized by extensive lateral accretion sets and scroll-bar topography, indicating the presence of point bar deposits and river with meandering planform. sandstone bodies of fluvial meandering, characterized by coarse-grained sandstone, good sorting, mainly trough cross-stratification (St), and the thickness of these sandstone bodies ranges from 1.5 -2 m and have lateral extension, about 150-200 m wide (Figs. 5.2 and 5.3a,b). The maximum thickness of the studied examples by

Gibling (2006), of meandering river deposits is 38 m, and widths are less than 15 km and typically less than 3 km.

5.2.3. Crevasse splay sand body

These relatively small channel deposits are present in ASF in two sheet of sandstones deposited by crevasse splay and consists of sandstone, fine to medium-grained, moderately to good sorting, and mainly current ripples and low angle cross-bedding, and from 0.3 - 0.5 m thick and about 20– 50m wide (Fig. 5.3a). Most crevasse splay bodies are less than 5 m thick and 50 m wide, with W/T commonly less than 10 (Gibling, 2006).

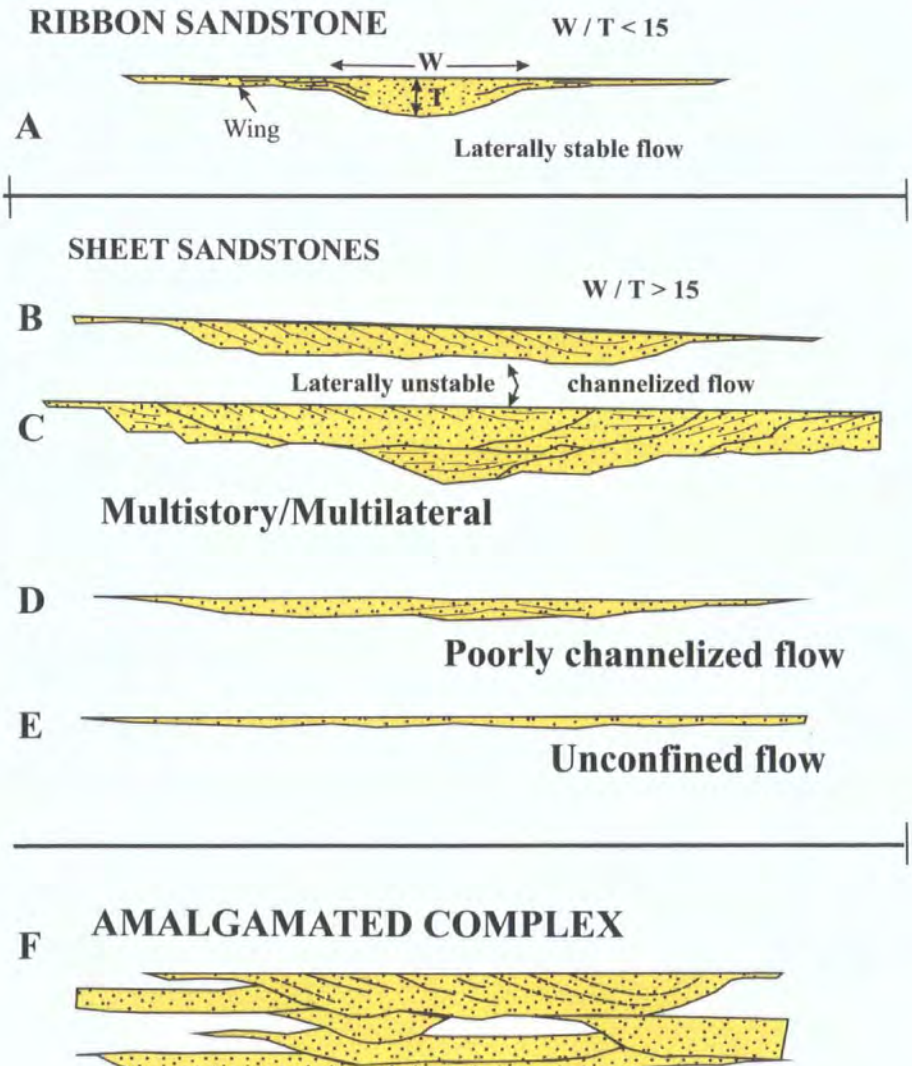


Fig. 5.1. Range of distinctive sandstone – body geometries recognized in the ASF (terminology as summarized in Hirst, 1983 ; Friend et al., 1986). A. Ribbon sandstone resulting from the plugging of a channel scour prior to any significant channel migration. Width / thickness ratio (W/T) measured perpendicular to pale flow are commonly between 5 and 10. The wings

attached to each side of the central body are coarse overbank deposits (levees). B to E Sheet sandstone bodies deposited by i. laterally unstable channels (B and C), ii. Poorly channellized flow (D), and iii. Unconfined overbank flow (E). Single sweeps of laterally unstable channels produce simple units (B) whereas migration of the channel back and forth results in more complex multi-storey / multilateral units (C). Unconfined overbank flow deposited units with very high W/T ratios and range from simple parallel-bedded tabular bodies to amalgamations of numerous small scour forms. F. An amalgamated complex of closely spaced sheet and ribbon sandstone bodies.

No	Lithofacies	Width (m)	Thick	W/T	No.	Lithofacies	Width (m)	Thick.	W/T
1	**CH**	50	3	17	**21**	**CH**	200	1.5	133
2	**CH**	60	2.5	24	**22**	**CH**	180	2	90
3	**CH**	50	1.25	33	**23**	**SS**	180	1.5	120
4	**CH**	45	1.5	30	**24**	**SS**	190	2	90
5	**CH**	40	2	20	**25**	**SS**	200	1.8	111
6	**CH**	55	1	55	**26**	**SS**	150	1.7	88
7	**SS**	60	1.25	48	**27**	**SS**	180	2	90
8	**SS**	25	0.45	56	**28**	**SS**	160	1.8	89
9	**CS**	20	0.3	67	**29**	**SS**	180	1.6	113
10	**CS**	50	0.5	100	**30**	**SS**	150	1.7	88
11	**SS**	40	1.7	24	**31**	**SS**	20	2	10
12	**SS**	80	2	40	**32**	**SS**	25	1.9	13
13	**SS**	70	1.5	47	**33**	**SS**	60	1.5	40
14	**SS**	90	2	45	**34**	**SS**	70	1.7	41
15	**SS**	85	1.5	57	**35**	**SS**	60	4	15
16	**SS**	190	2	95	**36**	**CH**	200	2.5	80
17	**SS**	200	1.5	133	**37**	**CH**	190	2	95
18	**SS**	190	2	95	**38**	**SS**	180	1.6	113
19	**SS**	200	1.5	133	**39**	**SS**	160	1.7	94
20	**CH**	180	2	90	**40**	**SS**	160	1.5	107

Table 5.1. Showing sandstone bodies' thickness and estimated width by meters selected from outcrop sections (Fig.. 3.1, sections 1, 2, 3 and 4) of studied area.

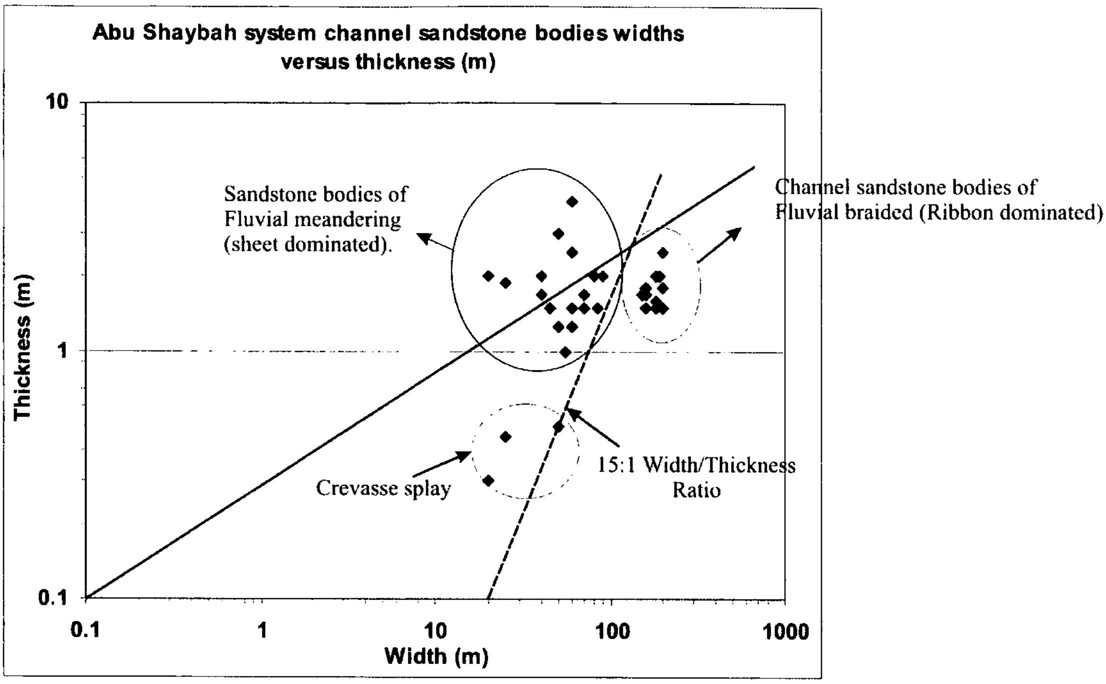


Fig. 5.2. Plot illustrating the width-thickness relationship (W/T) of channel-sandstone bodies at Wadi Ghan section (sheet dominated), and (ribbon dominated). The steeper line connects points with a 15:1 Width / Thickness ratio, which has been defined as the ratio separating ribbon from sheet sandstones (Friend et al., 1979).The widths of the units are measured perpendicular to paleoflow.

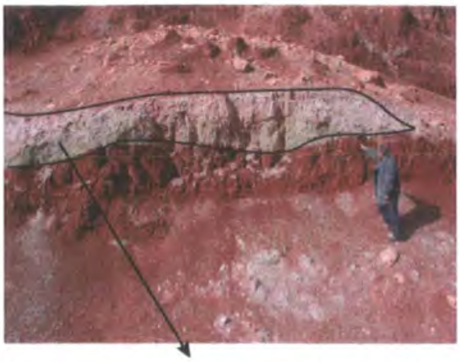
5.3. Importance of width / thickness for fluvial sandbodies

5.3.1. Valley fill W / T: Early stage of the ASF

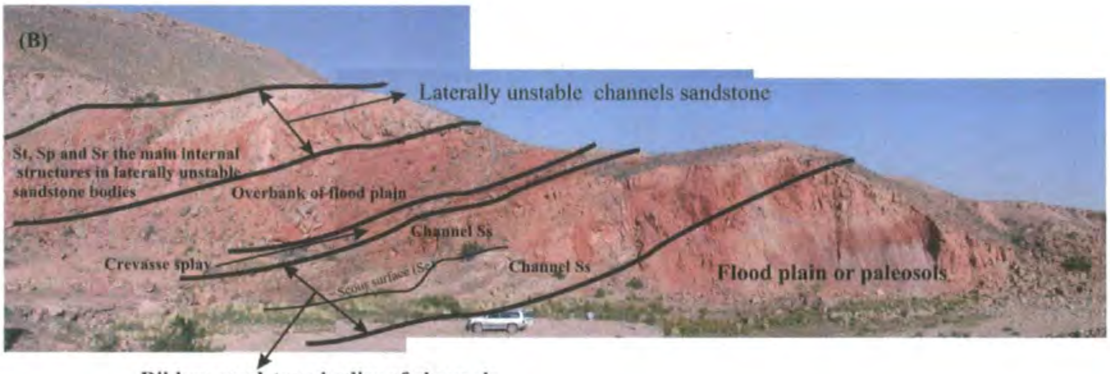
Valley fills have received considerable attention over the last decade from alluvial paleovalleys to incised submarine canyons (e.g. Vincent, 1999; Hampson *et al.*, 1999; Jones *et al.,* 2001; Jones, 2004; Feldman *et al.*, 2005). Although examples cover a large W/T space, they can be found more or less within the range of 2 to 100 and most tend to be less than 10, reflecting incision into the bedrock and fixing in position, often along structural lineaments.

The ASF is incised for the early phase of deposition and the facies dominated by braided fluvial systems. The early braided system as described in Chapter 3 highlights how the river and thickness represents a relatively long period of incision and focusing of the river channels, coincident with tectonic movement along localised faults. The predominantly valley fill is composed of gravelly sandstone, coarse to very coarse-grained sandstone, mud clasts scattered at the base and within cross-stratification, poorly sorted, and these show an enormous range of dimensions-from 0.5 -10 m thick , with W/T values from 40m to 60 m width of complex channels , with 17 – 55 m range, on the W/D ratios for the sand bodies in incised valley fill, reflecting incision into bedrock that resisted lateral plantation, as well as incision along structural lineation of Wadi Ghan fault zone (Fig..5.3b). and refer to relevant Fig. for this lower section only of the braided river. It needs to be just on the lower section as this is within the incised valley). The ASF palaeovalley fill is not on the same scale as many examples in the literature (e.g. Palaeogene Sis Conglomerate of Spain; Vincent, 1999), but it exhibits many of the characterizing criteria for their recognition. These include a basal unconformity, thick deposits, laterally confined, with complex structural and accommodation controls that are frequently locally fault-controlled (Fig.. 3.13, cross section of the ASF).

Climate may exert a strong control on palaeovalley form and the fill. Feldman et al. (2005) noted that valley fills differed during dry periods from those of wetter periods with the latter tending to be larger and more sand bodies evident with strong interconnectedness. In ASF section, it can be inferred that incision occurs during the most fluvial part of the climatic cycle, the sediment flux and discharge become effectively balanced as the climate becomes drier and, presumably, during this phase the river remains trapped within the previously incised valley system. As the climate begins to become humid, and stream



(A). Photographs showing the laterally unstable sandstone bodies in Abu Shaybah Formation, at Abu Ghaylan section.



(B). Ribbon and unstable channels sandstone bodies at the base of Abu Shaybah Formation, deposited by fluvial braided at Wadi Ghan section.

Fig. 5.3. Showing three type of the sandstones bodies and these distribution in ASF, (A and B) channel sandstone bodies of fluvial braided (Ribbon dominated), sandstone bodies of fluvial meandering (sheet dominated) and Crevasse splay sand body

discharge increases, the sediment flux is released and the river aggrades rapidly, infilling the incised valley until it breaks out onto the wider floodplain..

5.3.2. Changing aspect ratio for the ASF

To be able to explore the variability of the width/depth value for the ASF through the succession a histogram has been constructed for measurements of width/ maximum depth for the readily identifiable channels (Fig.. 5.4). The data used in this histogram are measured exact thickness and estimated width for single bodies, width and thickness ranges alone

The histogram indicates that a few channels with width/depth values of less than 4 are narrow sheets; most channels have values of 4 to 7 (Fig. 4.5). It is readily apparent from the histogram (Fig. 5.4) that channel dimensions change through time, with initial incision and confinement to a palaeovalley that widens through time with less lateral constraints and avulsion eventually through bank failure. Interestingly, Rodolfo (1989) recorded a width depth range of 2-5 for lahar-cut channels triggered by intense rainstorms and similar size values have been recorded by volcanic processes and rapidly filled (e.g. Fisher, 1977).

For the channel bodies in ASF with a given aspect ratio, the balance between bank migration rate and channels aggradation rate will determine the general channel-body geometry (Fig. 5.1). Because channel bodies are sensitive to minor sediment flux variations channels are likely to widen and show lateral accretion. The lateral accretion rate is only apparent in the ASF for the middle to upper part of the succession as when channels are laterally fixed in position, vertical accretion and a more dynamic system is likely to prevail. Reoccupation of preexisting channels is most likely to be a common process with in laterally confined systems and a composite channel body will result. However, this is not widely apparent in the ASF and may reflect a high sediment supply with a declining accommodation space and/or waning tectonic activity (Fig. 5.1).

The avulsion periodicity represents the formation of new channels on an unconfined plain, and leads to channel expansion and possible amalgamation (Fig. 3.13 and cross section of ASF). It has been noted that some channel bodies with low W/T may represent repeated reoccupation of long-lived drainage patterns during inactive periods (Allen, 1983). This is a frequently overlooked observation of low W/D systems and low net to gross reservoirs. Such a reoccupation of a channel system may result in regional well-connected channel systems and at out crop scale are not observable.

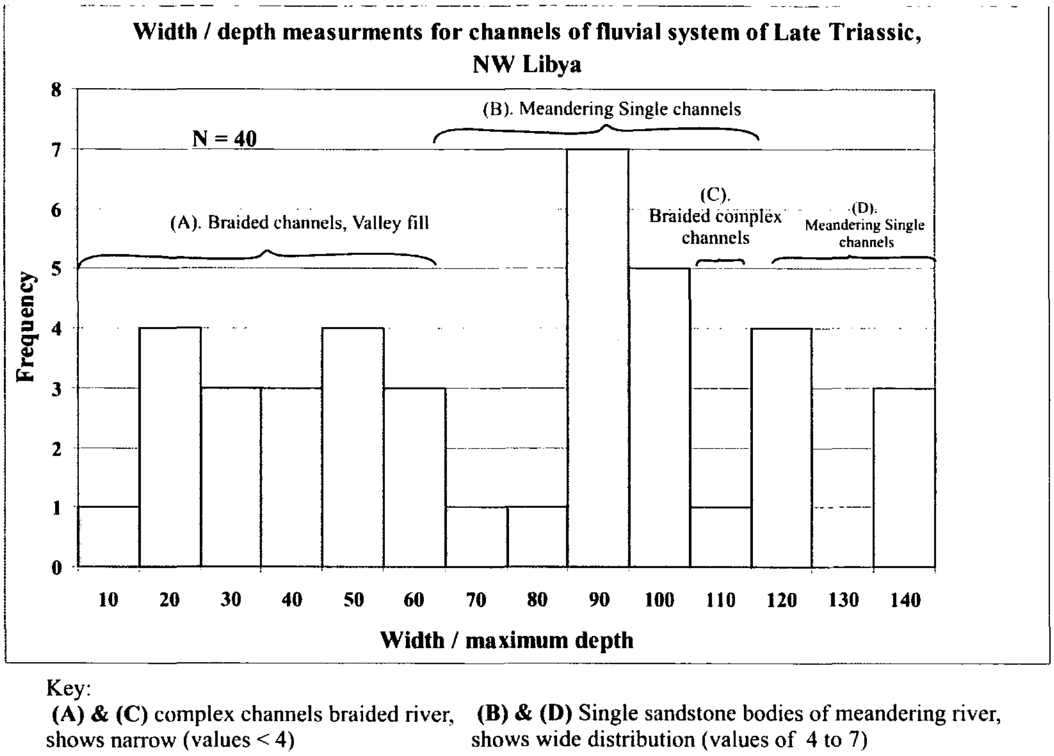


Fig.5.4. Width/depth measurements for channels of fluvial system of Late Triassic, NW Libya.

5.3.3. Channel bodies with low W / T

The ASF yields several channel bodies with exceptionally low W/T values (<3 narrow ribbons) (A & C, Fig. 5.4). Very low aspect ratios have been attributed to more than are factor: deep incision, high bank cohesive strength, rapid channel aggradation within short-lived channels, and reoccupation of previous channels (Allen *et al.,* 1983; Nichols, 1987; Smith, 1994).

The initial incised valley system of the ASF became fixed in position and the base is marked by a bedrock unconformity overlying the Al Aziziyah Formation. As a result the aspect ratios are perhaps less than to be expected with W / T of 15 or less. Similar W/T's have also been documented along faults where relatively narrow valleys are created (e.g. Carter, 2003; Fielding *et al.*, 2005). The low aspect ratio is due to the lateral confinement of the channels against bedrock and perhaps even syntectonic movement. The channels are stacked and although no criteria have been established it seems reasonable that channel reoccupation is a common process for the stacking of channel bodies. Furthermore the channel body thickness may exceed width under compactional conditions (e.g. Sander, 1989), but this has not been widely observed in the ASF.

5.4. Implications for reservoir modelling

The data acquired for the ASF presents several implications for modelling of fluvial systems.

I) Model results for thick and extensive formations that comprise fixed channel sand bodies (low W/T) are unlikely to contribute significantly to basin fills with high degrees of interconnectedness. These incised valleys, with stacked sand bodies of low W/T are not simulated in many models (e.g. Bridge, 1999) and freely meandering systems, where no bedrock control exists are frequently assumed. However, minor changes in gradient of the basin floor, older indurated alluvium, consolidated floodplain mud will all restrict the channel enlargement and behaviour of the river.

II) Interestingly most fluvial models are avulsion based, linked directly to elevation of the channel belt and its natural levees for avulsion to take place (Jones and Schumm, 1999). However, the ASF and the meandering deposits probably experienced avulsion due to increased sediment flux and high bank strength (even bedrock) promoted aggradation. Several authors have commented that many channels do not have levees or are only poorly developed (Hirst, 1991). In such cases, the choice for a new channel route may have little relationship to the basin floor gradient and can be 'randomly' placed.

III) A point of particular interest from the ASF is the absence of thick, extensive meandering sand bodies. Although theory would suggest meandering stream systems can generate, multistory sand bodies with a high degree of interconnectedness, there appears little indication that the ASF has done so. In contrast the braided rivers of the ASF have frequently created multistory deposits, which in part may be due to lateral constraints of the incised valley basement and structure.

IV) The study of the ASF suggests that for fixed channel sand bodies many additional factors come into play such as bank strength and local channel aggradation, as explored in the model of Makaske (1998) for anastomosing systems. The results of many modelling experiments suggest that fluvial architecture is also controlled by avulsion frequency, sedimentation rate and width of channel belt relative to the width of the basin (e.g. Heller & Paola, 1996). However, it is apparent from this study that once a channel becomes fixed then many of the models become inaccurate at predicting fluvial architectural patterns and ultimately reservoir geometries. Care must also be exercised when the channel sand bodies change from fixed braided to less restricted meandering streams, which may have been subject to different controlling factors, including climate change (e.g. Jones *et al.*, 2001).

Many of the model based studies have widely discussed assumptions and limitations associated with the models generated and most researchers have striven to attain the most realistic results from data sets available, whether subsurface or outcrop analogues. The study of the ASF offers in a similar way the ability to generate more widely applicable models. By combining the architectural models as outlined above with broader basinal models (Heller & Paola, 1996; Paola, 2000), through the use of sediment flux equations a more unified approach will be achieved especially between the interplay of fluvial architecture and external controlling factors such as climate and tectonics.

5.5. Sequence stratigraphy as applied to the ASF

A correlation between fluvial architectural patterns and systems tracts (Van Wagoner et al., 1990), more recently attempts have been made to relate variations parts of a valley-fill succession to lowstand (LST), transgressive (TST) and high-stand (HST) system tracts (Shanley and McCabe, 1994; Vail et al., 1991; Wright and Marriott,1993; Zaitlin et al., 1994). Publications on this subject note that many Holocene and ancient valley fills have a some what similar succession of deposits that grade upward from coarse or point bar deposits encased in extensive floodplain fines (Figs. 5.5 and 5.6).

Some authors suggest that the succession may be capped by extensive fluvial sheet sands. Explanations for this succession usually centre on the occurrence of higher gradients following relative sea –level, low- stand and incision, and changes in the rate at which accommodation is created throughout the filling and overtopping of valleys.

The lowest alluvium in many modern valley fills is coarse-grained and probably was deposited by a high-gradient low-sinuosity channel. These deposits generally change upward to deposits formed in large meandering streams and finally to those of smaller, modern, meandering channels (Schumm and Brakenridge, 1987).

The Late Triassic of ASF NW Libya displays a similar overall history in an area where tectonics and sea level change were negligible and climate change from relatively dry to wet to intermediate was probably responsible for the changing channel patterns.

The currently steps of this chapter will focus on application of sequence stratigraphy of ASF of fluvial system and incised valley fill to similar model of (Shanley and McCabe, 1994) and comparison to the one important subsurface reservoir rocks in middle to late Triassic fluvial system in North Africa TAG-I, Berkine basin, Algeria.

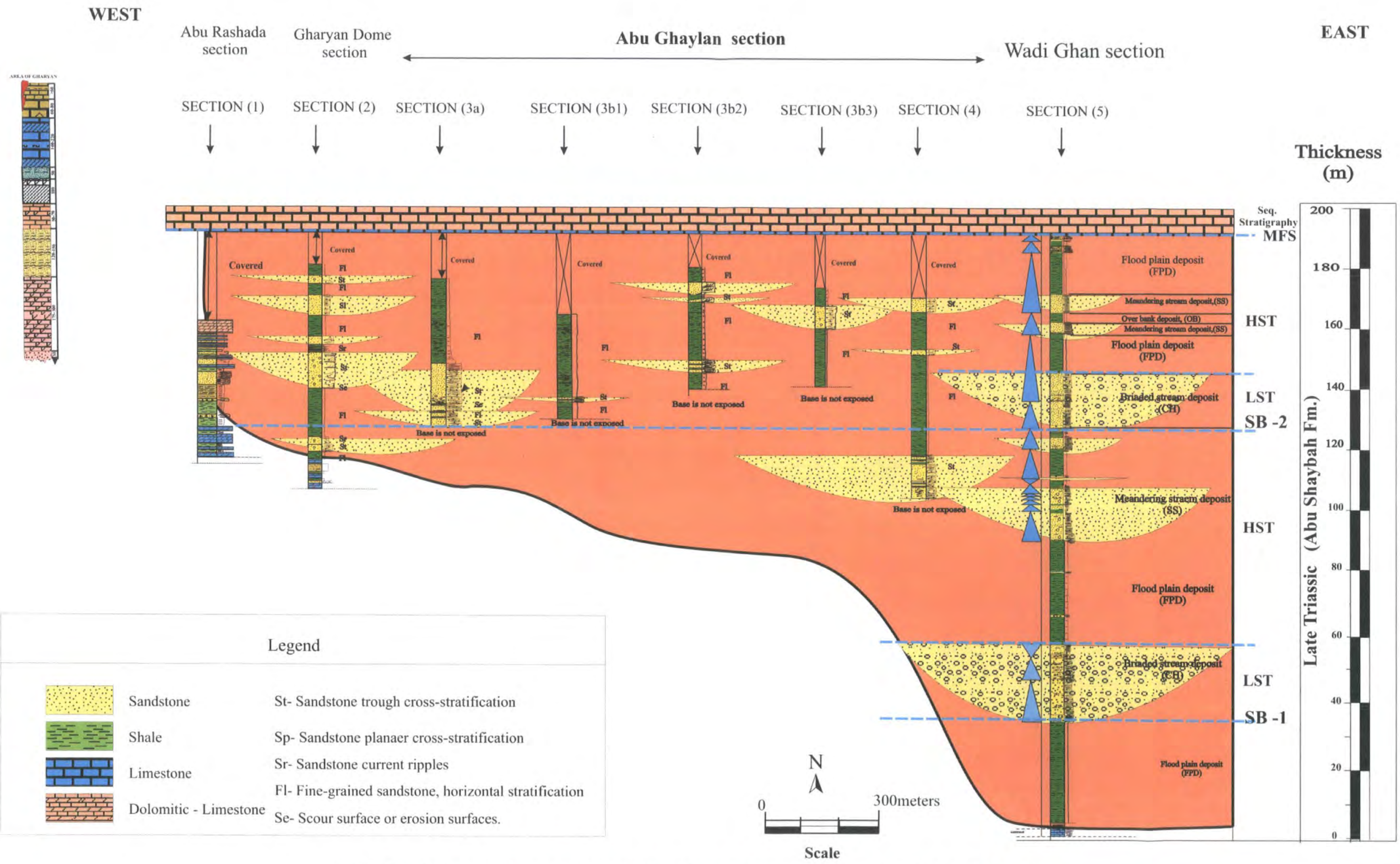
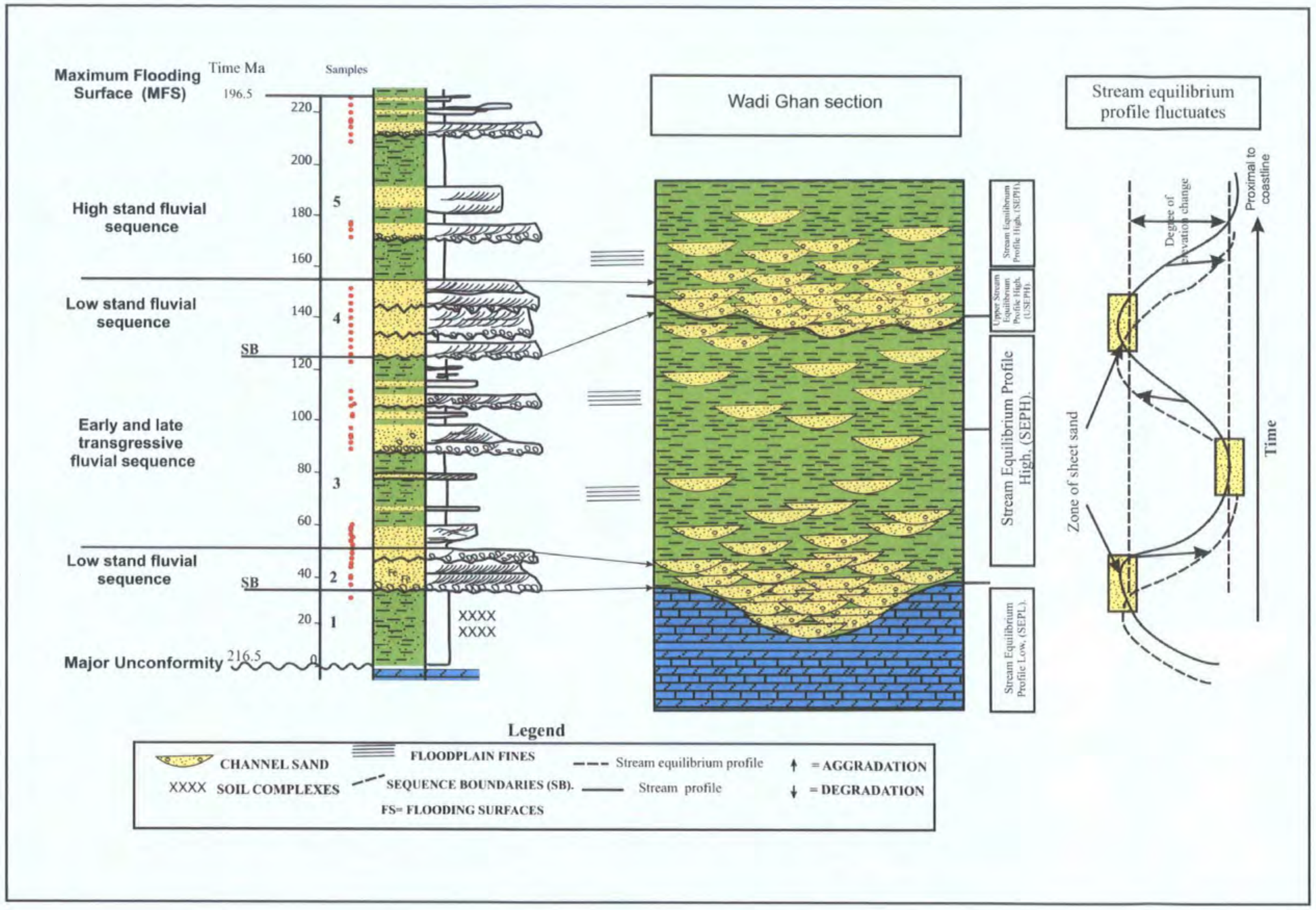
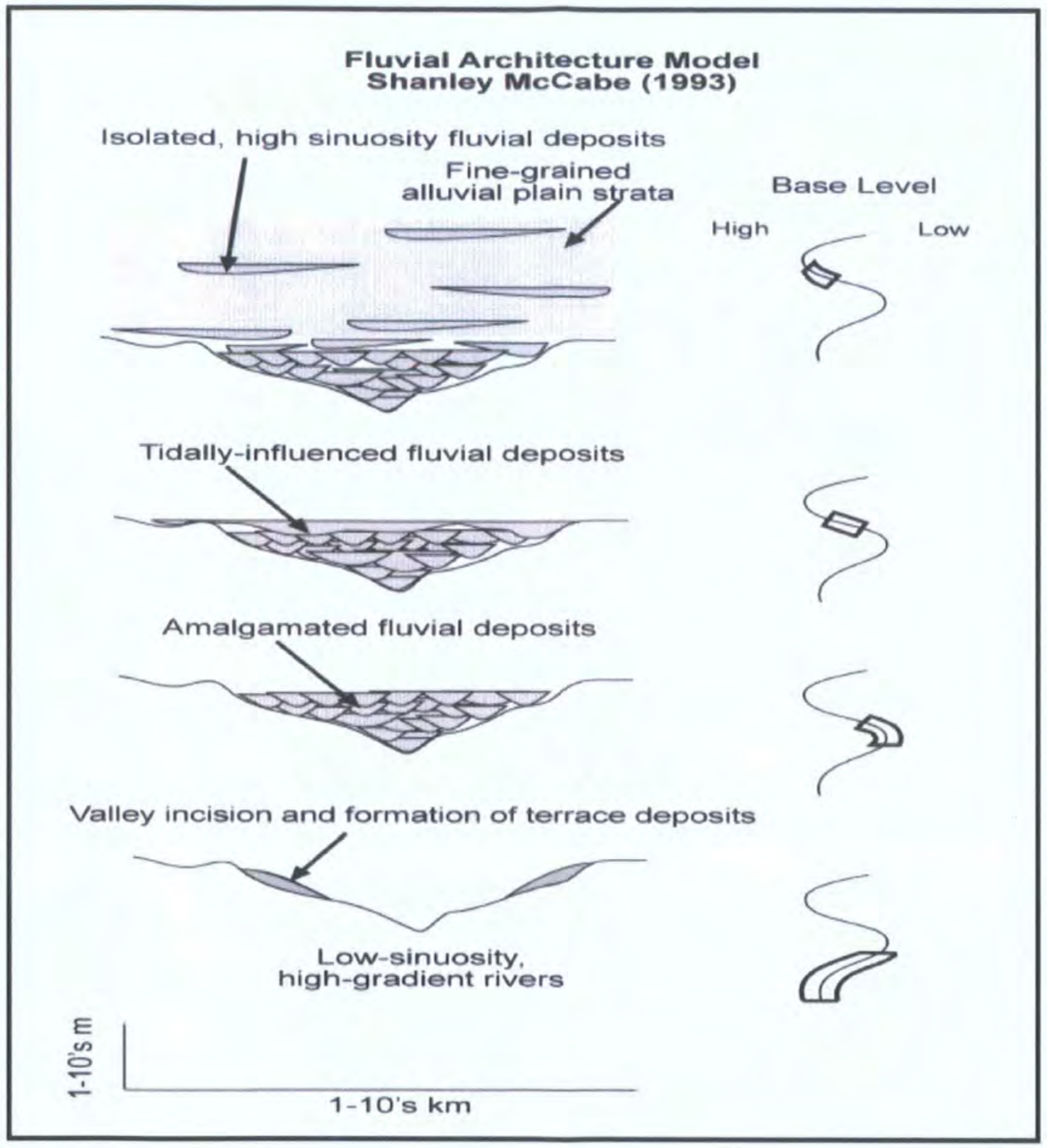


Fig.5.5. Lithological correlation panel along strike of the northern margin of Jabal Nafusah in ASF sections, showing two sequence boundaries started by Lowstand system tract (LST).



Abu Shaybah Fm. (Mayouf, 2007)



(Shanley and McCabe, 1994)

Fig.5.6. Schematic figure of the ASF illustrating the sequence stratigraphy on concepts of (Shanley and McCabe, 1994).

5.5.1. Lowstand Systems Tract (LST)

During periods of lowstand it is quite common for non-marine environments to experience sediment bypassing, erosion and incised valley development. The deposition of sediment during this time appears to be restricted to higher-accommodation environments where preservation is more probable. This is a situation in which tectonic subsidence will play an important role in the preservation of such deposits.

The actual outcome is dependant on the rate to which base level will create incision into the underlying strata. In theory, this incision of the incised valley should approach the depth to which the base level has fallen. Valley incision is not limited to eustatic sea level change but it may also be initiated from a change in the sediment load, water discharge or tectonic uplift of the source area (Richards, 1996). For deposition to occur in the lowstand of sea level, there needs to be deceleration of the abrupt fall in sea level (after the inflection point on the eustatic curve), allowing aggradation of sediments adjacent to incised valleys as this time is marked by sequence boundaries in the form of soil development. Terraces may also form on these sequence boundaries (Fig. 5.5).

The ASF of Wadi Ghan section is similar to the Blackhawk Formation, USA, which is an example of a fluvial lowstand systems tract with incision into the underlying sequence boundary, (major unconformity). Channels have a high degree of interconnectiveness and the sand: shale ratio is high as fine grained deposits are often cannibalised due to the low accommodation. Van Wagoner (1991) suggests that the multi-storey, multilateral Castlegate Sandstone of the Blackhawk Formation was deposited in the lowstand system tract. This sandstone was restricted in an incised valley that was 100 Km wide. He also points out that the Statfjord Formation shows the same fluvial architecture for the lowstand systems tract.

It is still debatable whether incised valley fill is part of the lowstand or transgressive systems tract. Basically it depends on whether the incised valley is large or small in size in relation to the sediment supply. If an incised valley is deeply incised and large, like incision into a steep shelf during a rapid fall in sea level, then it may take both the transgressive and highstand periods to fill the whole valley. If the valley is shallow and small, like a valley incising into sandy substrate and on a low angle shelf, then the valley could be adequately filled during the lowstand or the lowstand and subsequent transgression (Boyd and Diessel, 1995b, Fig. 5.7).

5.5.2. Transgressive Systems Tract (TST)

When base level rises and accommodation increases, incised valleys are flooded with the increase in accommodation initiating deposition of transgressive valley fill. Depending on the balance between the rate of base-level change, subsidence and sediments supply, the valleys would either become an estuarine environment, marine mud filled as in Abu Rashada section Fig. 3.1, section 6, Facies association 4, Figs. 3.9 & 3.10 , or they attract thick fluvial deposits with a tidal influence.

Fluvial aggradation and progradation will occur after the rate of base-level rise decreases, allowing the graded stream profile to readjust. As discussed above, the effect on fluvial architecture here varies with the rate to which base level is rising or filling. Vertical aggradation of fluvial deposits occurs when base-level rise is rapid, producing strata that are fine-grained, floodplain dominated, with isolated, high sinuosity channel deposits which are interpreted as meandering to anastomosing (Fig. 5.7) (Boyd and Diessel, 1995b).

In the model proposed by Shanley and McCabe (1993), shown in Figs. 5.5 and 5.6, the multi-storey, multilateral channel sandstones are deposited in the initial period of the transgression, and the tidally-influenced channels with flaser and wavy bedding, and dolomitic beds interlayered with fine sandstone and reddish mudstone with horizontal burrows occur at the maximum flooding surface. This surface separates the transgressive systems tract from the overlying highstand system tract. It is not until the late transgressive systems tract that the channels become isolated and switch from braided stream to meandering styles. Shanley and McCabe (1991b) also observed that the isolated channel sandstone beds are thinner in comparison to the amalgamated lowstand channel deposits. The early highstand systems tract shows the most isolated channels, which also illustrate the highest sinuosity. Boyd and Diessel (1995b) argue that the most isolated channels with high sinuosity occur at the maximum accommodation in the transgressive system tract, whereas the highstand systems tract deposits contain channel sandstones with higher degrees of interconnectiveness.

Caution is needed, as there are a number of inconsistencies to do with the response of non-marine sedimentation to a transgressive base-level cycle,

(Fig.5.7), shows channel deposits amalgamated with a high degree of connectiveness, which is interpreted as a transgressive systems tract, with the tidally influenced channels deposited during the maximum rate of transgression. The early highstand systems tract shows isolated channel deposits occurring when base level rise has slowed down. But one would expect that the channel bodies would be isolated in aggrading floodplain deposits (Fig. 5.6) when accommodation is at its maximum during the transgressive systems tract (Boyd and Diessel, 1995b). This is supported by the observations made by Van Wagoner (1991) who in the Blackhawk and the Statfjord Formations interprets the isolated, single storey sandstone channels, encompassed by floodplain mudstones to be from the transgressive systems tract.

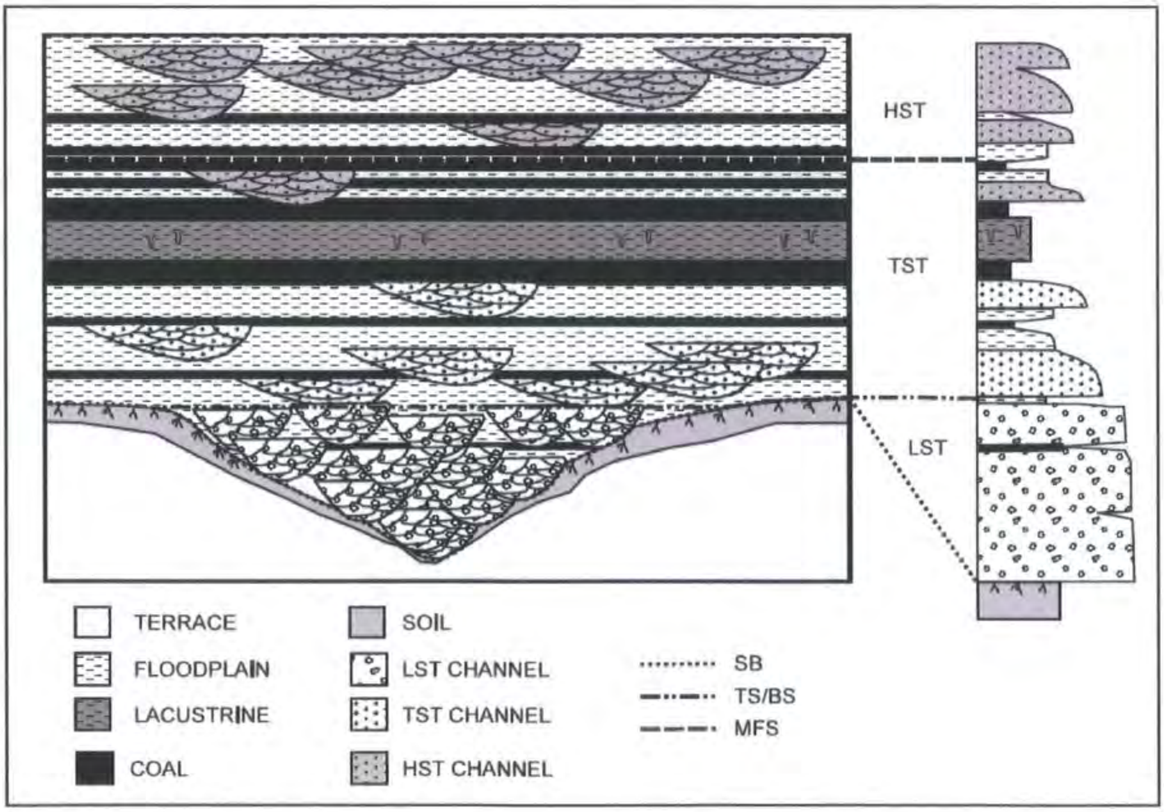


Fig. 5.7. Non – marine sequence stratigraphic model of Boyd and Diessel (1995b)

The Westphalian B Coal Measures of Northumberland and the southern North Sea also show a change in fluvial architecture with the change in the base level. With the increase in accommodation, the fluvial style change from multi-storey, multilateral sandstone to more isolated, thinner sandstones contained in fine-grained floodplain sediment (O'Mara and Turner, 1999).

5.5.3. Highstand Systems Tract (HST)

The base – level rate of change during the deposition of the highstand systems tract sediments is decelerating to a standstill and then begins to fall slowly. The highstand system tract is bounded by the maximum flooding surface at the base and by a sequence boundary at its top. Shorelines begin to regress under the slow rise of the base level and parasequences that were once retrogradationally are now aggradationally to progradationally stacked. Accommodation on the floodplain decreases, resulting in more lateral migration than vertical accretion of channel deposits, which creates an increase in channel interconnectiveness and amalgamation but not to the same degree that is found in lowstand fluvial deposits (Richards, 1996).

The highstand system tracts of fluvial deposits are often poorly preserved due to the oncoming lowstand, as there is a high possibility of these deposits being eroded. Either way, highstand systems tract deposits are generally thinner and sandier than strata that develop during the transgressive systems tract due to sediment bypassing the coastal area (Posamentier and Vail, 1988) and also due to the progressive decrease in accommodation. Van Wagoner (1995) found the highstand system tract of the Statfjord Formation to contain crevasse splays and levee deposits with low sand – shale ratio, but higher than what is observed in transgressive deposits.

5.6. Comparison to other Triassic reservoir sequence of North Africa (Fig. 5.8)

5.6.1. Algeria (TAG-I)

The Triassic Argilo-Gréseux Inférieur Formation (TAG-I) Middle to Late Triassic is one of the principal hydrocarbon reservoirs in the Berkine Basin of Algeria. Sedimentological studies have shown that it exhibits marked spatial and temporal facies variations on both a local field scale and a regional basinal scale. This variability, combined with a lack of diagnostic flora and fauna, makes regional correlation within the unit difficult. In turn, the lack of a consistent

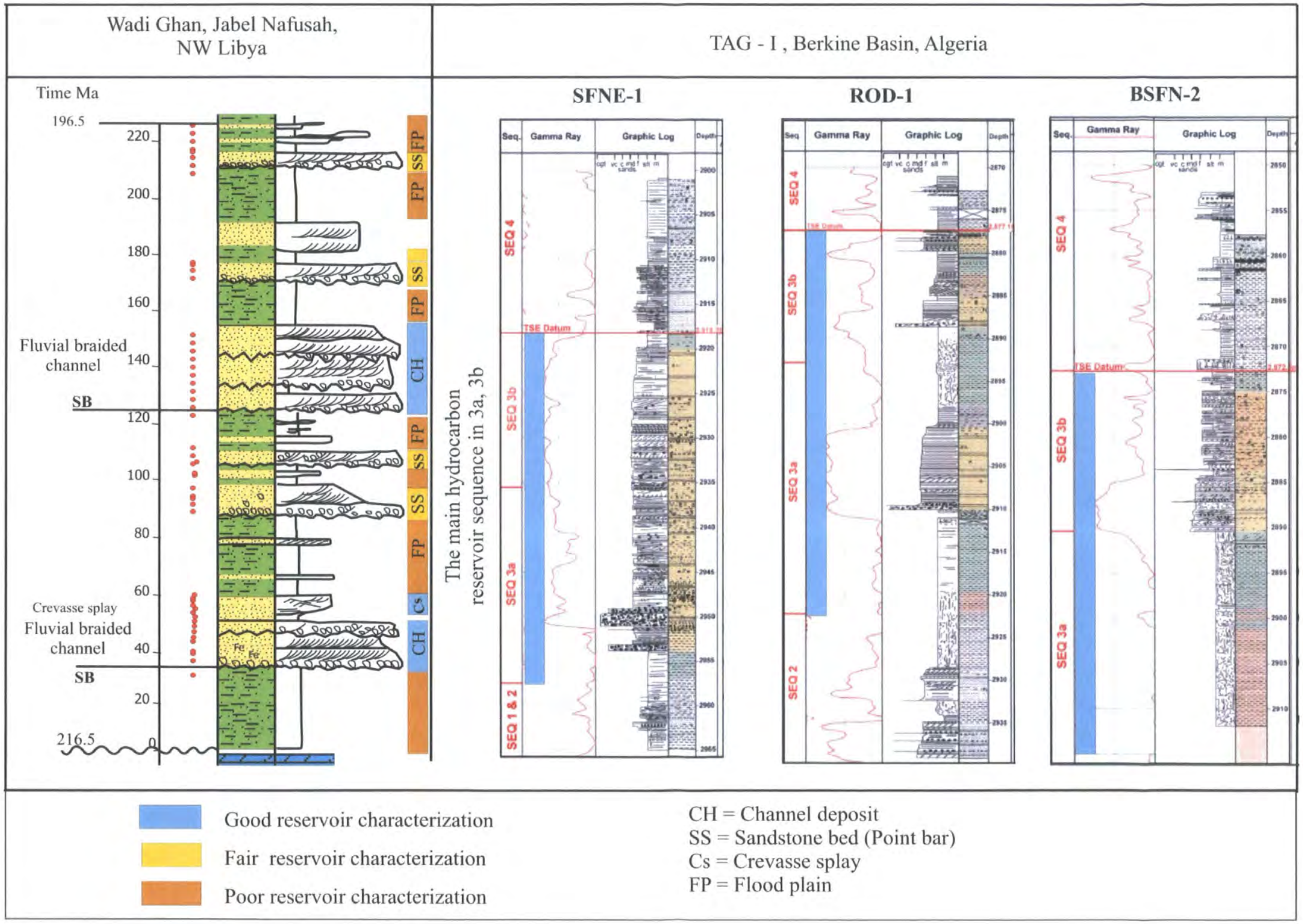


Fig.5. 8. Correlation of fluvial sequence architectural elements between ASF NW Libya and TAG-I, Algeria.

regional stratigraphic framework hampers the comparison of the various correlation schemes devised by operators in the basin.

The Triassic TAG-I (Trias Argilo-Greseux Inferieur) fluvial sandstones are the main oil reservoirs in the Berkine Basin, Algeria is not time equivalent and very similar unit to Late Triassic Abu Shaybah Formation in Jabal Nafusah, NW Libya and it also forms a major reservoir interval. The following explanation of diagenetic features and their effect on reservoir equality of the Abu Shaybah Formation (Late Triassic), TAG-I (Middle to Late Triassic) in Berkine Basin, Algeria is based on many references (Turner et al., 2001; Rossi, 2002; Pearce, 2004) state that the principal hydrocarbon reservoir is the Upper Triassic Argilo-Greseux Inferieur (TAG-I) which, in Blocks 401a and 402a, is Carnian to Norian in age. Elsewhere in North Africa, especially to the south and east, there are older Triassic Formations. For Example, in Zarzaitine in the extreme south east of the Triassic Algerian outcrop, Anisian vertebrates are documented (Fig. 5.8).

The TAG-I sits unconformably on Palaeozoic basement rocks with the basal Lower Carbonate comprises a laterally and vertically variable sequence which has been subdivided into four depositional sequences: Sequence 1, an unconformity-bounded, ephemeral fluvial intervals that fills palaeorelief on the Hercynian unconformity; Sequence 2, an initially upward fining, and subsequently upward-coarsening package of perennial fluvial sandstones and flood basin shales with thin crevasse splay elements and interfluve palaeosols; Sequence 3a,b, is distinguished by basin-wide fluvial incision and widespread channel sand deposition; Sequence 4, (Lower Carbonate) is a coastal plain and shallow marine system comprising shale, sabkha-type evaporates and bay-fill sandstones.

The Lower TAG I sandstones have an average framework composition of $Q_{98.3}F_{0.6}R_{1.1}$ and 95% of the quartz grains are monocrystalline. By contrast, the Middle-Upper TAGI sandstones have an average framework composition of $Q_{88.3}F_{9.8}R_{1.9}$ and 79.7% of the quartz grains are monocrystalline. The Lower TAG I quartz arenites derived from Paleozoic siliciclastics rocks, whereas the Middle-Upper TAG I subarkose originated mainly from metamorphic terrains. This change in provenance is a potential criterion for correlation within the TAG I. Also, this change has contributed to the significantly different diagenetic paths followed by the Lower TAG I quartz arenites and the Middle-Upper TAGI subarkoses. Grain-coating illitic clays are abundant in the Lower TAG I, where they exert a critical control on reservoir quality. Authigenic vermicular kaolinite largely consists of dickite that replaced previously formed kaolinite. Dickitization was followed by late-stage illitization related to the dissolution of detrital and authigenic K-feldspar. Quartz, the most abundant cement, was mainly sourced by the pressure- or clay-induced dissolution of detrital quartz and is a critical factor controlling the reservoir quality. Overall, quartz cement is more abundant in the Lower TAG I than in the Middle-Upper TAG I, and this increase correlates with a decrease in average porosity. Within the Lower TAG I, quartz cement abundance is stratigraphically very variable, which is in part related to facies controlled variations in grain-coating clay, resulting in major vertical variations in reservoir quality. Anhydrite and barite cements postdate quartz overgrowth. The sulfate necessary for their formation was likely sourced by deep subsurface dissolution of Late Triassic-Liassic evaporites.

The four depositional sequences reflect differences in depositional style resulting from base level shifts, tectonics and climate throughout TAG-I times, the overall increase in relative sea level was interrupted by periods of incision, which may relate to periods of rifting and erosion of the rift shoulders of the Berkine Basin. The initial valley (Seq. 1) was deposited under relatively arid or semi-arid condition. During seq. 2 and 3a, perennial fluvial systems with anastomosed channels and floodplain lakes became dominant and the climate increasingly humid. The common pore types in the ASF are intergranular and secondary porosities (grain dissolution, cement dissolution) and microporosity. Porosities in ASF ranges from fair to very good (0.2% to 23%) permeability was ignored. This porosity data was derived from point counting only (microscope study). And this variation in porosity in ASF is strongly related to facies variation, which was influenced by depositional environments and diagenetic processes. Good to very good porosity (23%) in the channels sandstone deposited by fluvial meandering, and very poor to fair porosity (0.25%) in the floodplain deposit. Cementation by quartz overgrowth, carbonate cements (calcite, dolomite and anhydrite) are the main carbonate cements and has great reduced porosity and authigenic clays pore lining chlorite and pore filling kaolinite are the effected in the pore through between the grains.

(1). TAG – I, Demonstrates the difficulty in correlating well logs in fluvial sequence and the ASF further demonstrates the complexities in field analogues.

(2). Important reservoir intervals of the TAG-I are characterized by basal erosional surfaces, typically where the fluvial system has incised LST. This surface can be the basis for correlation and are keys for understanding the evolution of a fluvial system.

(3). The ASF exhibits succession characteristics with incised valley fills and basal erosional. The type sequence of braided stacked channel passing up into floodplain and meandering stream, point bar deposits is regionally correlatable and can be classified as a Sequence unit LST – TST – LST.

5.7. Summary comments

Despite the obvious importance for subsurface applications, the three-dimensional geometry of fluvial sand bodies and valley fills has received relatively little attention since Friend (1983). This is surprising considering the wide use of width (W) and thicknesses (T) of channel sand bodies in reservoir studies and modelling. Furthermore there has been a tendency to use W/T and general channel sand body geometry to explain stratigraphic variations in terms of base level and accommodation space changes. Such a sequence stratigraphic approach to fluvial successions can be flawed as local geomorphological factors often determine channel forms, especially with incised valley fills such as the ASF.

The ASF has provided an improved understanding of channel bodies from laterally constrained valley fills to less confined mobile channel belts for the Late Triassic of NW Libya. A summary of the key findings are explained below:

1) The ASF is incised for the early phase of deposition and the facies dominated by braided river deposits. Faulting controlled the position of the valley fill and resulted in lateral confinement of the early fluvial system and resulting in lower W/T than would normally be expected (~40-60 m).
2) A changing aspect ratio is apparent for the ASF from laterally confined channels for the valley fill to more mobile channel belts in the middle to upper part of the Formation (Fig. 5.4). However, some channel bodies with low W/T may represent repeated reoccupation of long-lived drainage patterns. Such reoccupation may result in regional well-connected channel systems.
3) The W/T observations from the ASF have important implications for reservoir modelling. Two particularly important aspects are i) fixed channel sand bodies with low W/T are unlikely to contribute significantly to basin fills with high degrees of interconntedness and, ii) although theory and modelling would suggest that meandering stream systems generate, multistory sand bodies with high degrees of interconnectedness, there appears to be little indication that the ASF has done so, unless they have occupied previous channel systems.
4) The ASF represents a lowstand systems tract (LST) with incision into the underlying sequence boundary and low W/T and a high degree of

interconnectedness. This passes into a transgressive systems tract (TST) with a rising base level and dominated by isolated, high sinuosity meandering channel deposits. They have a moderate W/T ratio but are low net to gross systems with poor connectivity except where channel reoccupation has occurred and developed an established drainage system. The succeeding highstand systems tract (HST) records diminishing accommodation space and moderate W/T. Some interconnectedness is apparent but dominated by meandering stream deposits before the final marine flooding surface and Sequence boundary.

5) Comparison of the ASF with the TAG-I sandstones of Algeria illustrates many close similarities. The sequence stratigraphy applied to the ASF demonstrates that LST braided fluvial channel bodies are good exploration targets but in the ASF have low W/T. However, meandering stream channel sand bodies can also provide equally good reservoirs especially where channel reoccupation is widely apparent and develop into well-connected fluvial systems.

Chapter 6
Conclusions

Chapter 6

Conclusions

6.1. Key Aspects of Research:

The Late Triassic Abu Shaybah Formation (ASF) has been studied in outcrop localities along the Jabal Nafusah Escarpment, NW Libya. Through detailed graphic logging, petrography and architectural analysis a far greater understanding of the ASF fluvial systems has been obtained.

This research presents one of the first detailed sedimentological studies of the Late Triassic ASF. On first impressions this would only add to the already sizeable literature base on Triassic fluvial systems form North Africa, Iberia and the North Sea, UK. However, the sedimentological studies of the ASF fluvial system have demonstrated that the ASF is Late Triassic whereas much of North Africa consists dominantly of marine facies at this time; the low net to gross fluvial system of the ASF provides excellent exposures for W/T of channel sand bodies from valley fill to mobile channel belts; the study has illustrated, the role climate and tectonics have to play in controlling the evolution and behaviour of fluvial systems along the Jabal Nafusah and; detailed petrography of the fluvial channel sand bodies has recognised the importance of diagenetic mineral clay mineral growth for identifying sequence boundaries in fluvial successions such as the ASF.

The key conclusions from this study include:

1. Much of the Triassic fluvial deposition across North Africa occurred during the Early Triassic. The ASF in comparison is Late Triassic in age but contains very similar facies to many other Triassic petroleum plays of North Africa (e.g. TAG-I, Algeria). This provides potential scope for exploration into Late Triassic sediments further to the East of the Field area into the Al Kufrah and Sit Basins of Libya.
2. The ASF is located in an incised valley fill cited along a strike slip fault zone. Fluvial systems of the ASF were strongly controlled by strike slip movements into the faults currently Paleocurrent patterns and overall drainage.
3. The architecture of ASF of fluvial systems changed character and became less influenced by tectonism reflecting fewer multistory/ amalgamated fluvial deposits and more isolated channel networks while the climate in the fluvial sediments of ASF are marked by variable stream discharge and

sediment flux, in conjunction with ultimate base level, channel aggradation and degradation rate of fluvial systems.

4. Linking diagenesis of kaolinite and carbonate cements of ASF to depositional facies and sequence stratigraphy can provide a better understanding of some of the parameters that control the spatial and temporal distribution of diagenetic alterations and of their impact on reservoir quality. A study of fluvial system of ASF (Late Triassic of Jifarah Basin)

5. The ASF is incised for the early phase of deposition, and the facies dominated by braided river deposits under faulting, controlled in the position. The observation of changing ratios of W/T for low net to gross systems of ASF are apparent iṅ that the sandstone bodies of ASF concentrated the laterally confined channels in the incised valley fill and Mobil. Channel belts in the middle to upper part of the formation. This observation has important implications for reservoir modelling.

6.2. Suggestions for future work

During the process of this research, a number of unanswered questions have arisen concerning Triassic fluvial systems and the wider scope of fluvial architecture and reservoir modeling. This provides extensive scope for further research and with the recent and on-going research into Triassic reservoirs of the North Sea and Central Europe there is an ever-growing need for improved reservoir models from out crop analogues. Some suggestions for continued research include:

1) Improved characterization of fluvial architectural analysis through the use of terrestrial laser scanning, (ground based LIDAR studies), digital photogrammetry, high precision GPS survey etc.
2) Further detailed W/T studies of Triassic outcrops and reservoirs for a better appreciation of channel sand bodies during a time of rifting and high sediment flux.
3) Detailed appraisal of well-logs from the Sirte and Al Kufrah Basins, especially the deeper portions of the basins, for recognition of Late Triassic fluvial deposits and then potential as petroleum reservoirs.
4) Improved understanding of climatic indicators controlling fluvial depositional processes at the outcrop and basinal scales.

5) A better understanding of the transition from fluvial to marine facies for the Late Triassic of Libya, but also applicable to Buntsandstein to Muschelkalk facies across Europe.
6) Using the data acquired in this study to produce industry standard Petrel™ or GoCAD™ models of fluvial systems. The models should be created using a geomorological approach rather than the previous bias towards base level and tectonic controls.
7) The lack of understanding of mudstones is particularly apparent in fluvial successions where most research and economic interests are concerned with sandstone architecture, depositional patterns and allocyclic controls on fluvial channel sedimentation. Further research is required especially where thick floodplain fines exist and could be used for correlation in fluvial successions, especially in low net to gross systems.

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APPENDIX:

PETROGRAPHIC DATA SUMMARY CHART

PETROGRAPHIC DATA SUMMARY CHART

WADI GHAN SECTION

GENERAL											ROCK FABRIC (%)					MACR. TYPE PORES > 16µm		GRAINS (%)						AUTH. CLAYS (%)		AUTHIG. CEMENTS (%)					
SAMPLE No.	SAMPLE NAME	FORMATION	ENVIROMENT	FACIES (Ref. to Chapter 2)	CLASSIFICATION (Pettijohn, 1975)	GRAIN SIZE RANGE	SORTING	ROUNDNESS	GRAIN CONTACTS	STRUCTURE IN THIN SECTION	TOTAL GRAINS	TOTAL DETRITAL MUD	TOTAL CLAY	TOTAL CEMENT	MACROPORES > 16 µm	INTERGRANULAR POROSITY (%)	SECONDARY POROSITY (%)	MONO. QUARTZ	POLY. QUARTZ	LITHIC FRAGMENTS	FELDSPAR	MICA	HEAVY MINERALS	CHLORITE	KAOLIN	MUD MATRIX	QUARTZ OVERGROWTHS	ANHYDRITE	CALCITE	DOLOMITE	HEMATITE
1	WG-1	AZIZIYAH																													
2	WG-2	AZIZIYAH			Dolomitic limestone, Composed mainly Echinoderms, "Crinoids " and rare alge, coated by anhydrite.																										
3	WG-3	AZIZIYAH																													
4	WG-4	Abu Shaybah	FB	CH	Mudstone																										
5	WG-5	Abu Shaybah	FB	CH	QUARTZ ARENITE	F-M	MP	SA-SR		MA-G	71.9		15.9	2.5	9	2.8	6.2	66	2.2	0	2.6	1.1	Tr	6.5	9.4		2.5				
6	WG-6	Abu Shaybah	FB	CH	SUBLITHICARENITE	F-M	MG	A-SR	ST-CC	MA	74.1		6.5	6.5	12.4	7.4	5	62.1	6.2	2.5	2.2	1		2	4.5		6.5				
7	WG-7	Abu Shaybah	FB	CH	SUBLITHICARENITE	F-M	VP	A-SA	PO-GS	MA	61.9		7.3	30.6				52	6.3	1	2.6		Tr		7.3		1			9.6	20
8	WG-8	Abu Shaybah	FB	CH	SUBLITHICARENITE	M-C	MP	A-SR	ST-CC	MA	79.8		6.9	1.3	11.3	5	6.3	65.3	8	2.3	3.6	0.6	Tr	0.3	6.6		1.3				
9	WG-9	Abu Shaybah			MISSING																										
10	WG-10	Abu Shaybah	FB	CH	LITHICARENITE	MD	VP	A-SA	C.SUP.	MA	72.9		1.3	25.3				51	11.3	4.6	6				1.3		0.3			12.5	12.8
11	WG-11	Abu Shaybah	FB	CH	LITHICARENITE	F-VC	VP	SA-SR		DS	67.7			32.2				45.7	14	8							0.8			21.4	10
12	WG-12	Abu Shaybah	FB	CH	SUBLITHICARENITE						71.3	2.3	11	6.3	9.3	0.3	9	60	6	4.3		1	Tr	9	2	2.3	3				3.3
13	WG-13	Abu Shaybah	FB	CH	SUBLITHICARENITE						81.9		109	1.6	5	1	4	71	9	1.3	0.3	0.3	Tr	10.3	0.6		1.6				
14	WG-14	Abu Shaybah	FB	CH	SUBLITHICARENITE						58.6	2.6	6.6	26.2	5.6	2	3.6	54.6		3	1			6.6		2.6	0.3		3.6	22.3	
15	WG-15	Abu Shaybah	FB	CH	SUBLITHICARENITE						59.6	0.3		33.6	5.9	3.3	2.6	50.3	4	5.3						0.3	1				32.6
16	WG-16	Abu Shaybah	FB	CH	QUARTZ ARENITE						47.9		6.9	8.9	8.6	3	5.6	47.3		0.6				6.3	0.6		1.6				7.3
17	WG-17	Abu Shaybah	FB	CH	SUBLITHICARENITE	F-FU	MD	A-SR	PO	MA	64.8					0.2	1.4	58.7	6	0.5	0.5	0.5		2.5	1.1		2.2			22	4.8
18	WG-18	Abu Shaybah	FB	CH	SUBLITHICARENITE						55.9	3.3	9	19.9	12.6	2.3	10.3	50.6	1	4	0.3			9		3.3	3.6		7.3	9	
19	WG-19	Abu Shaybah	FB	CH	SUBLITHICARENITE						54	11	15	6.3	15.6	2	13.6	44.6	2.5	3	3			15		11	3		2	1.3	
20	WG-20	Abu Shaybah	FM	SS	SUBLITHICARENITE	VF-F	PO	A-SR	PO	MA	66.7		4.5	38.4				61.7	4.5	0.5				1.1	3.4		2	4.5	14.8	7.1	
21	WG-21	Abu Shaybah	FM	SS	QUARTZ ARENITE						70.2	1.6	12.6	3	12.3	2.3	10	68	Tr	1		0.6	0.6	11.6	1	1.6	3				
22	WG-22	Abu Shaybah	FM		MISSING																										
23	WG-23	Abu Shaybah	FM	SS	QUARTZ ARENITE						50.2	4	37		7.6		7.6	49.6	Tr			0.3	0.3	37		4	1				
24	WG-24	Abu Shaybah	FM	SS	QUARTZ ARENITE						85.8		7.3	2	4.9	0.3	4.6	85	0.5	0.3				7	0.3		2				
25	WG-25	Abu Shaybah	FM	SS							40	28.6	30	1.3				40						30		28.6	1	0.3			
26	WG-26	Abu Shaybah	FM	SS	QUARTZ ARENITE						71.9	2	9.6	6.6	9.8	7.6	2	69.6	Tr	2.3				9.6		2	6.6				
27	WG-27	Abu Shaybah	FM	SS	QUARTZ ARENITE						46	4.6	43.6	1	3.9	2.6	1.3	46						40.6	3	4.6	1				
28	WG-28	Abu Shaybah	FM	SS	SUBLITHICARENITE	F-M	MP	A-SR	ST-CC	L	74.4		3.6	5.7	13.7	5.7	8	68.8	0.5	1.4		1.7	2.2	2.2	1.4		5.7				
29	WG-29	Abu Shaybah	FM	SS	SUBLITHICARENITE	M-C	GD	SA-R	PO-CC	MA	51.4		19.9	4	23.9	13.1	10.8	46.5	0.8	2.2	1.7	0.2		18.5	1.4		4			Tr	
30	WG-30	Abu Shaybah	FM	FPD	SUBARKOSE	F-M	MG	SA-R	ST-CC	MA	63.1		13.9	7.1		7.1	8.2	55.1	0.5	2.2	3.7	0.8	0.8	8.5	5.4		7.1				

PETROGRAPHIC DATA SUMMARY CHART

WADI GHAN SECTION

GENERAL											ROCK FABRIC (%)					MACR. TYPE PORES > 16µm		GRAINS (%)						AUTH. CLAYS (%)		AUTHIG. CEMENTS (%)					
SAMPLE No.	SAMPLE NAME	FORMATION	ENVIROMENT	FACIES	CLASSIFICATION (Pettijohn, 1975)	GRAIN SIZE RANGE	SORTING	ROUNDNESS	GRAIN CONTACTS	STRUCTURE IN THIN SECTION	TOTAL GRAINS	TOTAL DETRITAL MUD	TOTAL AUTHIOGENIC CLAY	TOTAL CEMENT	MACROPORES > 16 µm	INTERGRANULAR POROSITY (%)	SECONDARY POROSITY (%)	MONOCRYSTALINE QUARTZ	POLYCRYSTALINE QUARTZ	LITHIC FRAGMENTS	FELDSPAR	MICA	HEAVY MINERALS	CHLORITE	KAOLIN	MUD MATRIX	QUARTZ OVERGROWTHS	ANHYDRITE	CALCITE	DOLOMITE	HEMATITE
31	WG-31	Abu Shaybah	FM	FPD	MISSING																										
32	WG-32	Abu Shaybah	FM	FPD	MUDSTONE					**Mainly Chlorite**														MAINLY CHLORITE							
33	WG-33	Abu Shaybah	FM	FPD	MISSING																										
34	WG-34	Abu Shaybah	FM	SS	LITHIC WACKE	F-M					61.7	24	3.3	3.4	7.1	1.4	5.7	55.1	0.2	1.7	1.7	2.8	0.2	1.1	2.2	24	3.4				
35	WG-35	Abu Shaybah	FM	SS	QUARTZ ARENITE	F-M	MG		ST-CC	L	64.7	0.8	17.6	4.5	11	4.5	6.5	63.7	Tr		Tr	Tr	1	4.5	13.1	0.8	4.5				
36	WG-36	Abu Shaybah	FM	SS	SUBLITHICARENITE	F-M	MG		ST-CC	MA	77.1	1.7	1.7	23.9	14.2	3.1	11.1	67.7	6.2	1.7	0.8	0.2	0.5		1.7	1.7	4.8				
37	WG-37	Abu Shaybah	FB	CH	QUARTZ ARENITE	Md	MP	SA-SR	ST-CC	MA	74.8	1		21.3				68.6	4	1.6			0.6			1	2.6	6.3	15		
38	WG-38	Abu Shaybah	FB	CH	SUBLITHICARENITE	M-C	Md	SR-R	PO-CC	MA	77.1	4.3	0.6	14.6	2.9	1.3	1.6	69	5.6	1.6	0.3		0.6	0.3	0.3	4.3	2.3		12.3		
39	WG-39	Abu Shaybah	FB	CH	QUARTZ ARENITE	Md	MG	SR-R	ST-CC	MA	79.3		1.6	3.9	14.9	9.6	5.3	74	4	1.3				0.6	1		3.3		0.6		
40	WG-40	Abu Shaybah	FB	CH	QUARTZ ARENITE	Md	MG	SR-R	ST-CC	MA	81.9	3	0.9	2	11.9	5.6	6.3	76.3	1.3	2	1	1	0.3	0.6	0.3	3	2				
41	WG-41	Abu Shaybah	FB	CH	SUBLITHICARENITE	F-M	MG	SR-R	ST-CC	MA	76.1		5.6	8.6	9.3	5.3	4	67.6	2.3	2.3	1.3	1	1.6	0.6	5		8.6				
42	WG-42	Abu Shaybah	FB	CH	SUBLITHICARENITE	C-VC	GD	0	CC	MA	72.6		1.7	22.7	1.6	1.1	0.5	64.8	7.7	0.1					1.7		1.7	4.8	12	4.2	
43	WG-43	Abu Shaybah	FB	CH	MISSING																										
44	WG-44	Abu Shaybah	FB	CH	SUBLITHICARENITE	Md	PO	SA-SR		SL	75.8		12.6	6.3		3.3	1.3	67	4.6	1.3	0.3	Tr	2.6	2.3	10.3		6.3				
45	WG-45	Abu Shaybah	FB	CH	MISSING																										
46	WG-46	Abu Shaybah	FB	CH	SUBLITHICARENITE	Md	MG	SA-R	PO-CC	MA	76.5		2.6	8		8		66.6	3.3	2	2	Tr	2.6		2.6		8				
47	WG-47	Abu Shaybah	FB	CH	SUBARKOSE	Md	Md	A-SR	ST-CC	MA	78.6		4.3	4		9.3	3.6	66.6	4	3	4		1		4.3		4				
48	WG-48	Abu Shaybah	FM	SS	LITRHERNITE	F-C	Vp	A-SR	GR.SUP	MA	75.5	5.4	11.6	2.5		3.1	1.7	51.7	18.8	3.6	0.6	0.8		7.1	4.5	5.4	2.5				
49	WG-49	Abu Shaybah	FM	SS	QUARTZ ARENITE	M-C	Md	SA-SR	ST-CC	MA	89.8		1	1.6		4	3.3	84.3	2.3	0.6	1.3		1.3		1		1.6				
50	WG-50	Abu Shaybah	FM	SS	SUBLITHICARENITE	Md	MG	SA-SR	ST-CC	MA	83.8		2.6	4.3			2.3	70	4.3	3.3	5.6		0.6		2.6		4.3				
51	WG-51	Abu Shaybah	FM	SS	SUBLITHICARENITE	F-M	Md	A-SR	ST-CC	MA	69.9		18.6	2.3		5.3	3.6	60.3	3	3	3.3		0.3	16	2.6		2.3				
52	WG-52	Abu Shaybah	FM	SS	LITRHERNITE	C-VC	Vp	SR-WR	GR.SUP	MA	84.5		12.2	2		1		59.9	23.3	0.3		1		11.6	0.6		2				
53	WG-53	Abu Shaybah	FM	SS	QUARTZ ARENITE	F	Mp	SA-R		L	84.4		3.6	4.6		7	0.3	78.2	0.3	1	2.3	0.6	2	2	1.6		4.6				
54	WG-54	Abu Shaybah	FM	SS	SUBLITHICARENITE	F	MG	A-SR		L	76.4	1.1	12.2	3.1		3.1	3.7	62.8	0.8	1.4	3.7	4.5	3.2	11.7	0.5	1.1	3.1				
55	WG-55	Abu Shaybah	FM	SS	SUBLITHICARENITE	F-M	MG			MA	75.3		11.1	2.5		6.5	4.2	67.4	2	2.5	1.4	1	1	11.1			2.5				
56	WG-56	Abu Shaybah	FM	SS	SUBLITHICARENITE	Md	PO	SA-R	ST-CC	MA	81.6	2.3	1	3.6		6	5.3	73	6	1.6	Tr		1	1		2.3	3.6				
57	WG-57	Abu Shaybah	FM	SS	QUARTZ ARENITE						63.6	1.3	29.3	0.6		2.6	2.3	61	Tr	2	Tr	0.3	0.3	28	1.3	1.3	0.6				
58	WG-58	Abu Shaybah	FM	SS	SUBARKOSE	F	PO	A-SR	P-MS	MA	59.5	2	25	3.3			9	52	0.6	0.6	5	0.6	0.7	25		2	3.3				
59	WG-59	ABU GHYLAN			**Algae mat, ABOUT 30% OF BIVALVA FLOATING IN CALCAREOUS CARBONATE MUD.**																										
60	WG-60	ABU GHYLAN																													